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Multi-Sensor Analysis of Thunderstorm Microphysics and Electrical Response to Hailfall in the Metropolitan Area of Bologna

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Abstract

This work presents a multi-sensor diagnostic analysis aimed at characterizing the internal cloud microphysical structure and electrical response of intense convective storms to hail-fall events within the metropolitan area of Bologna. Specifically, the study investigates whether precipitating solid hydrometeors act as a physical vector for downward charge transport, effectively leading to the electrical discharge of the convective core. The observational framework combines ground-based measurements from a laser disdrometer with 3-D volumetric data from a polarimetric Doppler weather radar network, alongside a real-time lightning tracking network.

The analysis is applied to 17 hailfall events extracted from the disdrometer historical archive spanning from December 2018 to August 2025. On the microphysical scale, a detailed disaggregation of hydrometeor spectra from the total dataset of the hailfall events indicates a contamination of the liquid phase by solid particles. This highlights the limitations of the classification algorithm of the disdrometer employed in this study, rendering it less reliable, especially in mixed-phase conditions where melting occurs during fall. In parallel, the analysis of the total lightning dataset compiled for these events reveals a major anomaly. Contrary to the regional Mediterranean climatology, the total lightning distribution in these selected case studies is inverted, with cloud-to-ground discharges strongly dominating over intra-cloud pulses.

A detailed analysis of the temporal evolution of both disdrometer observations and lightning activity across the most intense case studies is conducted. Results show that the highest electrical activity occurs prior to the arrival of hail at the ground, followed by a progressive decline during convective core grounding and a near-complete electrical collapse once the hailfall ceases, indicating that solid hydrometeor fallout drives the discharge of the thundercloud.

The integration of polarimetric radar data enables the study to overcome the point-source limitations of ground-based instruments, providing a comprehensive view of the convective systems' spatial evolution and internal core dynamics. By cross-referencing instruments featuring different temporal resolutions, the combined analysis of lightning activity, 3-D volumetric radar data, and specific extracted parameters allows for the explanation and deeper understanding of anomalous signatures observed during the previous temporal analysis.

Ultimately, this integrated framework highlights the necessity of volumetric multi-parameter coupling to fully resolve the lifecycle, charge distribution geometries, and structural transitions of localized thunderstorms.

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Introduction

Understanding the coupling between the formation and growth of ice particles and electrical activity is crucial for monitoring convective storms and improving the knowledge of cloud structure and dynamics. This thesis investigates, from an experimental point of view, the electrical response of thunderclouds to solid precipitation, specifically focusing on the active role of hail as a downward charge carrier, driving the depolarization of the convective core. To this aim, an extensive dataset comprising 17 convective events recorded by a laser disdrometer is integrated with high-resolution lightning observations from a ground-based sensor network. In addition, data from ground-based polarimetric radars are utilized for the most relevant cases to investigate the three-dimensional microphysical structure of the storms. Measurements from these various platforms are carefully synchronized and combined, highlighting the close relationship between the cloud's electrical activity and the occurrence of hailfall.

Hail is produced within deep convective storms characterized by strong updraft, high cloud tops, intense lightning activity and a sufficient lifetime to support hydrometeor growth. Within the European continent, the Po Valley in northern Italy represents one of the most severely exposed and hazard-prone regions for such phenomena, regularly experiencing devastating economic impacts due to hailfall (Punge and Kunz, 2016) [32]. An in-depth understanding of these severe storms requires a detailed observation of the hydrometeors reaching the surface. From a microphysical perspective, the characterization of these convective events relies indeed on the analysis of the Drop Size Distribution (DSD), which quantifies the drop concentration per unit volume as a function of their diameter. How the DSD evolves across distinct precipitation regimes has been thoroughly investigated in the scientific literature. While stratiform precipitation is well-modeled by a simple inverse exponential distribution (Marshall and Palmer, 1948) [22], convective precipitation exhibits different signatures. In intense convective regimes, the DSD is indeed better represented by a three-parameter Gamma distribution (Ulbrich, 1983) [37], reflecting a much broader spectrum characterized by the presence of large raindrops. Under conditions of extreme precipitation, the DSD reaches an equilibrium determined by the balance between drop coalescence and their breakup. Consequently, the median volume diameter saturates around a constant value, and further intensity increases are driven by higher drop concentrations rather than growth in hydrometeor size (Bringi et al., 2003) [6].

When intense convection triggers the production of solid precipitation, the ground-level Hailstone Size Distribution (HSD) broadens to include large particles. As these solid hydrometeors descend below the freezing level, the melting process causes smaller hailstones to melt completely before reaching the ground, depleting the concentration of

smaller particles and flattening the distribution at lower sizes (Fraile et al., 2003) [14].

The generation of solid hydrometeors and the macroscopic electrification of the thundercloud are physically intertwined. The very same intense updrafts required to suspend and grow massive ice particles also drive continuous collisions between ice crystals, super-cooled liquid water, and graupel within the storm’s mixed-phase region. This collisional mechanism represents the primary engine for cloud electrification.

While the microscopic mechanisms of electrification and charge accumulation within thunderclouds have been extensively investigated, through historical laboratory experiments (Takahashi, 1978) [34], detailed three-dimensional numerical simulations (Mansell et al., 2005) [21] and experimental observations (Roberto et al., 2016, Buiat et al., 2017) [33] [7], the inverse process of cloud depolarization remains largely unresolved. In particular, significant observational gaps still surround the electrical response of convective systems to the fallout of solid hydrometeors, such as hail and graupel, and their active role as physical downward charge carriers.

Historically, the few studies addressing this topic have primarily been dedicated to electric field measurements. During the Airborne Field Mill II (ABFM II) campaign, conducted between 2000 and 2001, *in situ* aircraft observations provided comprehensive datasets on microphysics and electric fields in the anvil region of the thunderstorm [12]. These observations were subsequently utilized by Willett and Dye (2003) [39] to develop a theoretical bulk conduction model for predicting electrical decay times, which was then compared against the actual electric field measurements recorded during the campaign (Dye et al., 2007) [12]. Conversely, ground-based observational studies during the storm’s dissipation stage have been exemplified by Pawar and Kamra (2013), who collected experimental measurements of surface electric field variations beneath decaying convective systems [27].

Consequently, high-resolution observations capturing the real-time depolarization of the active convective core during the actual grounding of solid hydrometeors remain highly limited. This thesis aims to address this specific gap by laying the groundwork for the investigation of the electrical response and microphysical transition of severe thunderstorms during hailfall events over the metropolitan area of Bologna. To achieve this, the present work proposes an integrated, multi-sensor diagnostic analysis that combines surface microphysical data from a laser disdrometer, 3-D volumetric structures from polarimetric weather radars, and high-resolution 3-D lightning localization provided by the LINET network.

The thesis is structured into five chapters as follows. The first establishes the theoretical framework, reviewing cloud microphysics, thunderstorm dynamics, electrification mechanisms, and lightning phenomenology. The experimental setup and the operating principles of the employed instruments are detailed in the second chapter. The third outlines the methodology developed to process, filter, and integrate these multi-source datasets, while the fourth presents the core results of the research, providing a statistical overview of the analyzed convective events and a detailed 3-D radar analysis of the most extreme hailstorms to investigate the microphysical-electrical coupling. The final fifth chapter summarizes the main conclusions of this work and outlines future research outlooks.

Chapter 1

Theoretical Framework

1.1 Cloud Microphysics and Precipitation

In this section the fundamental microphysical processes governing cloud and precipitation formation are outlined. Understanding these dynamics is essential to interpret the macroscopic signatures detected by meteorological instruments, such as weather radars and disdrometers. The following subsections are structured as follows: first, exploring the thermodynamic mechanisms of water vapor condensation and ice nucleation; second, describing the collisional interactions, such as coalescence, aggregation and riming, that drive the growth of hydrometeors; and finally, illustrating the genesis of precipitation reaching the ground.

1.1.1 Cloud Formation and Nucleation

Cloud droplet formation is governed by the saturation ratio, defined as $S = e/e_s$, where e represents the partial vapor pressure and e_s the saturation vapor pressure at ambient temperature. The nucleation process is classified as homogeneous if it occurs spontaneously within pure vapor, or heterogeneous if it is mediated by aerosol particles, known as Cloud Condensation Nuclei (CCN). Homogeneous nucleation requires a saturation ratio of 3 – 4, values that are not attained in the atmosphere. Actually, cloud droplets form exclusively through heterogeneous nucleation, where CCN, such as insoluble particles, soluble particles or ions, enable condensation at the low supersaturations typical of the atmosphere, exceeding unity by only a small fraction (e.g., $S \approx 1.001$ -1.01). Following nucleation, droplets increase their size through diffusional growth as water vapor molecules condense onto their surface. This process continues until the droplets reach a critical diameter where their terminal fall velocity becomes non-negligible, allowing the drops to interact with each other.

Homogeneous nucleation of ice directly from water vapor requires a saturation ratio of approximately 20, a condition never attained in the atmosphere. Conversely, homogeneous ice nucleation from liquid droplets occurs spontaneously at temperatures below -40°C . Such conditions are typically found in the upper regions of deep convective clouds, where temperatures can drop below -80°C . In the temperature range between -40°C and 0°C , liquid droplets are referred to as supercooled droplets. Given that ice formation occurs in natural clouds at temperatures above -40°C , an alternative mecha-

nism must exist to initiate ice formation, such as heterogeneous nucleation. This process is mediated by specialized particles known as Ice Nuclei (IN). IN with a crystal lattice structure similar to that of ice are the most effective at promoting nucleation and thus ice crystals themselves serve as the most efficient IN.

Following nucleation, ice crystals increase in mass through diffusional growth. The process is similar to the growth of liquid droplets, but the wide variety of ice crystal habits makes the process difficult to describe analytically. Once these ice particles reach a sufficient size, their terminal fall velocity becomes non-negligible, enabling the collisional interactions that govern the cloud’s microphysical structure.

1.1.2 Collisional Interactions

As a consequence of their growth, cloud droplets develop varying dimensions and terminal velocities, leading to interactions and collisional processes. Three distinct outcomes can result from these collisions. When two small droplets collide, they behave like rigid spheres due to high surface tension and bounce away. Conversely, when collisions involve larger droplets, the outcome depends closely on their sizes and impact energy: they either merge together to form a large drop through permanent coalescence, or they undergo temporary coalescence followed by separation, splitting into two or more fragments [20]. Collision-coalescence processes also occur among solid particles, but they are considerably more complex than for liquid water droplets, as the result of the interaction varies according to the type of the colliding particles. The collection mechanism for ice crystals, known as aggregation, governs the formation of snowflakes and strongly depends on their shape, temperature, and environmental conditions. Aggregation becomes more efficient when the temperature is in the range between $-1\text{ }^{\circ}\text{C}$ and $+1\text{ }^{\circ}\text{C}$; in this regime, the outer branches of the crystals undergo superficial melting, making it easier for them to bond with other particles. At lower temperatures, aggregation occurs through the interlocking of the branches, and dendritic crystals have a significantly higher probability of collecting other crystals.

When a supercooled droplet impacts an ice crystal, latent heat of fusion is released as the droplet freezes. This accretion process, known as riming, can produce different types of solid hydrometeors. If the latent heat is completely dissipated, the dry growth takes place where the droplet freezes almost instantaneously, producing a low-density (approximately 0.15 g/cm^3) agglomerate known as graupel. Conversely, if the latent heat is not entirely dissipated, a liquid water film remains on the surface, leading to the wet growth regime. This excess water can be aerodynamically eliminated during fall (shedding), resulting in the formation of a smooth ice layer, and the process is called glaze growth. Alternatively, the unfrozen liquid infiltrates the interstices of the ice, causing the hydrometeor to grow heterogeneously resulting in spongy growth. Both of these latter regimes lead to the formation of high-density hydrometeors (typically ranging from 0.8 to 0.9 g/cm^3), namely hail.

1.1.3 Precipitation Genesis

Once precipitating hydrometeors reach sufficient size, raindrop formation proceeds through two distinct mechanisms. In the first case, rain originates from the melting of ice particles

as they descend into atmospheric layers with temperatures above 0°C . While snowflakes melt rapidly and completely, graupel and hail can survive substantial fall distances, undergoing only partial melting. This mechanism typically dominates precipitation production in mid- and high-latitude regions. The second mechanism, characteristic of equatorial and tropical regions, is known as warm rain, where the ice phase is entirely absent, and raindrops grow exclusively through liquid-phase collision and coalescence processes.

When a raindrop reaches a dimension where internal surface tension can no longer counteract hydrodynamic instabilities and internal oscillations, it undergoes aerodynamic breakup. Theoretical and experimental studies pioneered by Pruppacher and Pitter (1971) [31] demonstrate that an isolated falling drop breaks up upon reaching a diameter of approximately 9 mm. In natural precipitation, however, raindrops rarely exceed a diameter of 6 mm. This restriction is the direct result of the continuous feedback between two competing mechanisms: coalescence and collisional breakup [20]. Collisional breakup occurs when the kinetic energy introduced by the impact between two drops cannot be dissipated through internal oscillations and viscous deformation of the merged entity, resulting in a small number of large remnant drops and a high number of very small droplets. Nevertheless, under exceptional convective conditions, such as in unpoluted marine clouds or intense tropical systems, "super-large" raindrops with diameters ranging from 8 to 10 mm have been observed [17].

1.2 Thunderstorm Dynamics

The formation of a thunderstorm begins with the genesis of a cumuliform cloud, typically triggered by surface solar heating. As solar radiation warms the Earth's surface, heat is transferred to the air layers above. This localized heating causes the less dense air to rise, expanding and cooling adiabatically as it ascends, until it reaches the Lifting Condensation Level (LCL) where water vapour condenses, leading to cloud droplet formation as described in Section 1.1.1. Under unstable and perturbed atmospheric conditions, this initial cumulus cloud grows vertically into a cumulonimbus, driven by a powerful current of ascending moist air. If the updraft is sufficiently strong, the cloud top reaches the freezing level, where temperatures drop below 0°C . Here, supercooled droplets and ice crystals collide, leading to the formation of graupel and hail. The vertical development of the cloud is halted by the tropopause, the stable layer typically located at 10 to 12 km at mid-latitudes, which forces the cloud to expand horizontally into its characteristic anvil shape. However, in extremely severe storms, violent updrafts can locally overcome this limit, creating an overshooting top.

This process, initiated by solar heating, is classified as a convective thunderstorm, but other external forcing generates the vertical motion. For example, orographic thunderstorms arise when air masses are forced to ascend along mountain slopes, while frontal thunderstorms develop when cold, dense air masses act as a wedge, forcing warmer air upward.

The most common type of convective system is the single-cell thunderstorm, typically triggered by small-scale gradients of temperature and humidity that create favorable conditions for convective initiation. Single-cell storms are generally small in spatial scale and characterized by a brief life cycle, lasting less than an hour. The life cycle of such

a convective cell evolves through three distinct phases. It begins with the developing stage, which is dominated entirely by the updraft transporting warm, moist air aloft. As liquid drops, graupel and hail accumulate sufficient mass to overcome the updraft's force, they begin to precipitate, initiating the mature stage. This falling precipitation generates a strong current of cold descending air, the downdraft, that coexists with the updraft. The mature stage typically lasts 10 to 15 minutes and is characterized by maximum storm intensity, heavy precipitation, and lightning activity. The last phase, the dissipating stage, begins when the cold downdraft spreads horizontally at the ground surface, cutting the inflow of warm, moist air. Deprived of the vertical inflow, precipitation decreases, and the cloud gradually evaporates. In some cases, the advancing edge of the downdraft spreading at the surface acts as a gust front, lifting the surrounding warm air and triggering the genesis of new cells. This mechanism leads to the formation of multicell systems, made up of individual cells in different phases of their evolution that regenerate and remain self-sustaining. These convective systems occupy vast areas and can last for several hours.

The most extreme convective phenomena are supercell thunderstorms. A supercell is characterized by an intense and rotating updraft known as the mesocyclone. This rotation is sustained by strong wind shear, which tilts the storm's structure and maintains the updraft and downdraft well separated, without interfering with each other. This spatial separation allows the storm to persist for several hours. As the rotating updraft extends toward the surface, it often manifests as a localized mesoscale low-pressure area, capable of generating strong winds. The descending currents frequently produce downbursts, powerful columns of descending air that strike the surface and spread out in all directions, causing severe wind gusts. These events are often associated with hailstorms, intense lightning activity, and sometimes tornado formation. From a radar perspective, supercells often exhibit a Bounded Weak Echo Region (BWER), which appears as a localized 'hole' in radar signal. This is caused by an intense updraft that transports particles so rapidly that they remain too small to be detected by radar until they reach higher altitudes.

1.3 Cloud Electrification

1.3.1 The Global Electrical Circuit

The Earth's atmosphere is subjected to a constant electrical influence that extends from the surface to the ionosphere. Measurements conducted under "fair-weather" conditions over the Earth's surface have confirmed the existence of an electric field, known as the Fair Weather Electric Field (FWEF), generated by the net negative charge residing on the Earth, which typically averages 130 V/m. This field is not static, instead it exhibits significant temporal and spatial variations, ranging from 50 to 300 V/m depending on local conditions, the time of day, and solar activity. The atmosphere contains both positive and negative ions, produced by natural radioactivity and cosmic rays. Under the influence of the FWEF, these ions are subjected to electric forces: positive ions are driven downward toward the surface, while negative ions migrate upward. This charge flow constitutes the atmospheric leakage current, which amounts to approximately 1500 A. Due to the conductivity of the atmosphere, if no external mechanisms were present to

replenish these charges, this current would effectively neutralize the FWEF in around 6.5 minutes. In reality, observations confirm that the FWEF remains persistent, implying the existence of a continuous charging process that constantly compensates for this leakage. This mechanism is driven by thunderstorm activity, and it is estimated that at any given moment, between 1500 and 2000 thunderstorms are active worldwide.

The Earth's electrical environment can be conceptualized as a massive spherical capacitor, where the conductive surface of the Earth and the ionosphere act as two concentric armatures, with thunderstorms operating in between as global charge generators. These storm cells counteract fair-weather dissipation by transferring positive charges toward the ionosphere, while global lightning activity acts as the primary impulsive mechanism that deposits net negative charge to the Earth's surface through cloud-to-ground discharges, thereby successfully maintaining the global circuit in a quasi-stationary state.

1.3.2 Charge Center Organization

A lightning discharge is a high-current atmospheric phenomenon that occurs when the localized electric field within a thunderstorm intensifies sufficiently to overcome the dielectric strength of the surrounding air, initiating an electrical breakdown, which will be described in detail in Section 1.4.2. This intense electric field within the thundercloud is macroscopically produced by the physical separation of large amounts of opposite electrical charges across vast volumes of the cloud, a process known as cloud electrification. Depending on the storm's convective intensity, its thermodynamic characteristics and the ambient vertical thermal profiles, these charge pockets arrange themselves into distinct geometric configurations. Four primary charge structures have been identified within convective cells. The standard configuration observed in convective clouds is the tripolar structure, illustrated in Figure 1.1. It consists of a main positive charge region located in the upper levels of the cloud and a main negative charge center located in the central region of the storm, complemented by a smaller positive charge layer situated near the cloud base. In contrast, during the earlier stages of a convective cell or within weaker convective environments, a simpler dipolar structure is frequently observed, which lacks the lower positive charge layer and exhibits a straightforward vertical separation consisting of an upper positive charge region and a lower negative charge center.

In certain convective environments, often associated with anomalous storms or intense hailstorms, the vertical layout of the polarities inverts completely. In these inverted structures, the negative charge center occupies the upper regions of the cloud, while the positive one resides at the low levels. Beyond these standard vertical configurations, convective systems often exhibit complex geometries where the charge distribution deviates from this vertical alignment. For instance, under the influence of strong environmental wind shear, the convective updraft can tilt, causing the positive and negative charge centers to displace horizontally and form an inclined bipolar structure, whereas other configurations are characterized by a broad horizontal distribution of alternating charge layers with minimal vertical development.

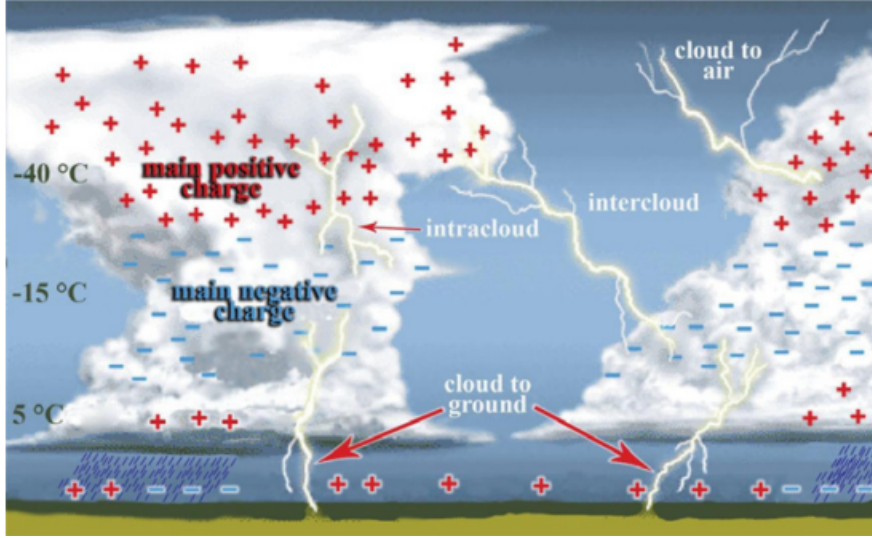


Figure 1.1: Representation of the electrical structure of a typical thundercloud, exhibiting a tripolar charge distribution. The primary lightning flash types – intra-cloud, cloud-to-ground, inter-cloud, and cloud-to-air – are illustrated. *Source:* Reproduced from Dwyer and Uman [11].

1.3.3 Inductive and Non-Inductive Mechanisms

In order to explain the macroscopic electrification of thunderclouds, several distinct physical mechanisms have been proposed, though determining which processes are effectively the most efficient remains a subject of ongoing research. There are two main classes of collisional charging mechanisms: inductive and non-inductive. Inductive, or polarization, charging requires a pre-existing electric field to induce charge on the surfaces of the colliding particles. Non-inductive mechanisms, on the other hand, operate without regard to an external electric field.

In particular, the inductive charging results from rebounding particle collisions in the presence of an electric field, which polarizes the particles by forcing ions of positive sign to accumulate at opposite sides. Consequently, charge may be transferred between the hydrometeors when they collide and separate. However, this polarization charging is ineffective for ice-ice collisions because the lower conductivity of ice and short contact times prevent electrical current from transferring appreciable charge [21]. To evaluate the actual efficiency of these mechanisms, Mansell et al. (2005) [21] conducted numerical sensitivity experiments using a three-dimensional cloud model simulating a multicell thunderstorm. Their results conclusively showed that when the simulation relied strictly on inductive charging, it failed to produce any strong electrification, leading to internal electric fields that were barely larger than the FWEF, thereby demonstrating that the inductive mechanism alone is insufficient to trigger initial lightning discharges. Instead, the non inductive riming electrification process acts as the primary driver of cloud electrification, being capable of initiating massive charge separation from an initially neutral state and generating the robust macroscopic electric fields required for lightning activity. A comprehensive description of this non-inductive mechanism will be detailed in the following subsection.

1.3.4 Riming Electrification

The quantitative framework of the non-inductive riming electrification mechanism was established through pioneering laboratory experiments conducted by Takahashi (1978) [34]. He studied collisional charging with a chamber containing ice crystals and super-cooled droplets, where rotating rods inside the apparatus acted like graupel particles by collecting droplets and colliding with crystals [21]. The empirical results extracted from these laboratory simulations demonstrated that both the sign and the order of magnitude of the electrical charge transferred to the graupel are strictly dependent upon the ambient temperature and cloud water content (CWC). At temperatures warmer than approximately -10°C , the rimed target systematically gains a net positive charge from rebounding collisions with ice crystals, regardless of the ambient cloud CWC regimes. At temperatures colder than approximately -10°C , the polarity of the acquired charge exhibits a more complex dependency on the CWC. The rime continues to gain a positive charge at both exceptionally low and exceptionally high extremes of CWC, but its magnitude remains too low to build the strong electrification required for typical lightning discharges. Instead, the hydrometeors become heavily electrified, acquiring a net negative charge, only at intermediate CWC values ranging between 1 and 2 g m^{-3} .

In the conclusions of his study, Takahashi (1978) [34] utilized these results to explain the formation of the tripolar structure (Figure 1.1). Within the convective updraft, as graupel forms and remains suspended in the colder regions of the cloud where temperatures are below -10°C , it systematically acquires a net negative charge through collisions with ice crystals, thereby establishing the cloud's main negative charge core, at altitudes between 6 and 8 km. When graupel falls into the region where the temperature exceeds -10°C , positive electrification of the graupel occurs and some graupel particles change sign from negative to positive, directly accounting for the generation of the positive charge region near the cloud base, at altitudes of about 1 to 2 km. The ice crystals acquire the compensating positive charge after their collision with graupel and due to their mass and lower terminal fall velocities compared to the graupel, these positive ice crystals are efficiently lofted by the updraft toward the highest altitudes of the convective system, typically near the tropopause at heights of approximately 10 km or above, establishing the upper positive charge center of the storm. Furthermore these elevated ice crystals are highly susceptible to upper-level wind shear, which displaces or tilts the upper positive core relative to the underlying charge centers, introducing the characteristic asymmetric layout observed in mature thunderstorms.

1.4 Lightning Phenomenology

A lightning discharge represents the transient and impulsive response of the atmospheric medium to the local breakdown of its dielectric strength. On a planetary scale, this phenomenon is remarkably frequent: global satellite and ground-based observations indicate that approximately 30 to 100 lightning flashes occur across the globe every single second, translating to nearly 9 million discharges per day that continuously maintain the Earth-ionosphere electrical balance.

Historically, the features of these discharges below the cloud base were first identified in the early twentieth century using rotating-lens optical systems, specifically the Boys

camera. In recent decades, these pioneering optical observations have been significantly enhanced by high-frequency digital video cameras and advanced electromagnetic interferometric techniques. Nowadays, modern total lightning detection networks allow scientists to reconstruct both the visible sub-cloud channel development and the hidden breakdown processes occurring deep inside the cloud cores by capturing radio frequency emissions and optical pulses.

1.4.1 Classification of Tropospheric Discharges

Following the taxonomy detailed by Dwyer and Uman (2014) [11], all atmospheric lightning discharges can be divided into two primary categories depending on whether they bridge the gap between the cloud charge centers and the Earth or remains confined within the atmosphere. The latter group, designated as cloud discharges, represents the statistical majority of all global lightning activity. Within this category, events occurring entirely within a single cloud are defined as intra-cloud (IC) lightning, which stands as the most common form of atmospheric discharge. Discharges occurring between separate clouds are classified as inter-cloud lightning, while those establishing a connection between a cloud charge region and the surrounding air are termed cloud-to-air lightning, as previously illustrated in Figure 1.1.

The other category includes cloud-to-ground (CG) flashes, which are further divided into four distinct types according to the sign of the electrical charge carried in the initial leader and the vertical direction of leader propagation. Approximately 90% of all ground discharges are negative CG flashes, which are initiated by a negatively-charged, downward-propagating leader originating from the negative charge core in the middle region of the cloud (Figure 1.2, top left). Downward positive CG flashes, initiated by a positively-charged leader moving downward from either the upper or lower positive charge centers, account for roughly 10% of ground events (Figure 1.2, bottom left). The remaining two configurations represent ground-to-cloud discharges that are relatively uncommon and typically originate from prominent features such as mountaintops, tall man-made towers, or other tall objects, propagating upward toward the cloud charge regions (Figure 1.2, top right and bottom right).

Over the last decades, numerous observational studies have been carried out to evaluate the numerical partitioning between IC and CG discharges. In one of the earliest global studies, Prentice and Mackerras (1977) [30] conducted an extensive literature review, aggregating 29 independent observations to measure the ratio between cloud flashes and cloud-to-ground flashes. A fundamental conclusion shared by these historical observations, performed through naked-eye visual counts and primitive field-change counters, is that this ratio systematically remains greater than unity, proving that IC flashes inherently outnumber CG events under standard planetary conditions.

More recently, advanced total lightning observations from the modern LINET network, collected through worldwide observation campaigns by Höller et al. (2009) [18], have provided a comprehensive cross-regional evaluation of these dynamics, confirming the global dominance of cloud discharges. However, they observed that the IC to CG ratio is dictated by local geographical regions, convective regimes and orographic forcing. Specifically, tropical, monsoon-driven continental systems exhibit the highest prevalence of cloud discharges, reaching IC fractions above 80%, whereas mid-latitude summer storms

or thunderstorms forced by mountain ranges show a significant increase in the cloud-to-ground fraction, with the IC contribution dropping below 70% [18].

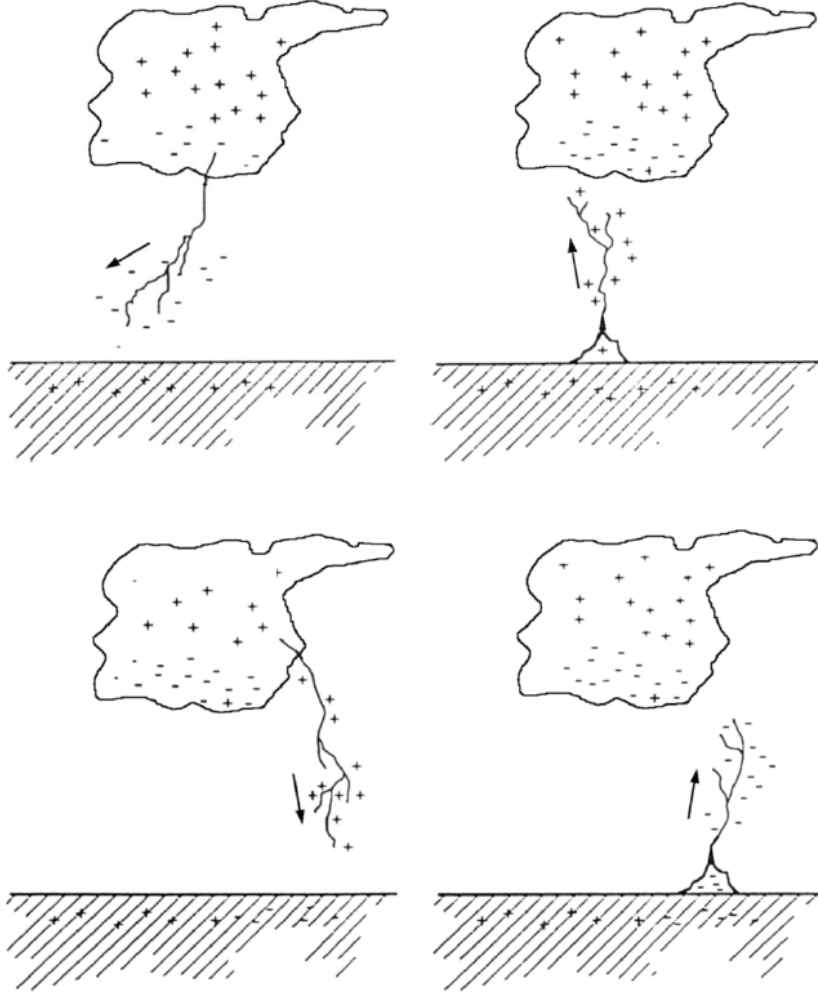


Figure 1.2: The four types of cloud-to-ground lightning flashes: (top left) the downward negatively-charged leader, (top right) the upward positively-charged leader, (bottom left) the downward positively-charged leader, and (bottom right) the upward negatively-charged leader. *Source:* Reproduced from Dwyer and Uman [11].

1.4.2 Lightning Discharge Mechanism

The physical process of a lightning discharge represents a complex sequence of electrical breakdowns, the mechanism of which is detailed below focusing on the Cloud-to-Ground flash as the primary model for sub-cloud electrical activity. The following description focuses specifically on the negative downward CG flash, which represents the vast majority of ground events, while the distinct characteristics of the positive CG flash are briefly summarized at the close of the section. The phenomenological description and quantitative values presented throughout this section adhere to the comprehensive framework established by Dwyer and Uman (2014) [11].

Furthermore, it is important to address the observational constraints associated with

Intra-Cloud (IC) lightning. While IC flashes are numerous within convective systems, taking place between the main positive and negative charge layers, they remain hidden from a structural standpoint. Because the entirety of an IC discharge occurs deep within the tropospheric cloud, its electrical mechanisms cannot be directly observed using photographic or optical analysis, meaning that all intermediate processes cannot be tracked or understood as clearly as they are in a CG flash.

Preliminary Breakdown and Stepped Leader Initiation

A negative Cloud-to-Ground lightning flash originates near the lower boundary of the negative cloud charge center. This initial phase, known as the preliminary breakdown, manifests as a series of radio frequency pulses or electric field deflections detected in association with the onset of the flash. For CG flashes, the preliminary breakdown essentially describes the internal Intra-Cloud lightning activity that occurs before the leader channel exits the cloud on its way to the ground. As the breakdown progresses, these intense preliminary pulses eventually disappear before the leader exits the cloud base, followed by a brief and quieter period characterized by only a few low-amplitude pulses. Although the cloud base is typically situated at an altitude of 1 to 2 km, this preliminary activity occurs much higher within the cloud, at altitudes of about 7 km. Consequently, it is impossible to observe the process directly and only the diffuse light emitted by the breakdown can be seen filtering through the cloud.

Structurally, this preliminary breakdown activity gives rise to a leader channel developing toward the Earth's surface as a stepped leader, as illustrated in Figure 1.3 at $t = 1.00$ ms. The stepped leader's movement from cloud to ground is not continuous, but rather it moves downward in intermittent, discrete segments of tens of meters in length (Figure 1.3 at $t = 1.20$ ms and $t = 19.00$ ms). Each individual step occurs at regular temporal intervals of about $50\ \mu\text{s}$ when the stepped leader is far above the ground, decreasing to approximately $10\ \mu\text{s}$ as it approaches the surface. A typical stepped leader has about 5 C of negative charge distributed over its length, carrying a charge density of about $10^{-3}\ \text{C m}^{-1}$ when approaching the ground boundary. In terms of channel geometry, the conducting core carrying this charge configuration is remarkably narrow, measuring only a few centimeters. In order to establish and maintain this highly charged electrical configuration during the downward descent, an average continuous background current of approximately 100 to 200 A must flow steadily through the channel throughout the whole leader process. However, every time the leader propagates forward by a single step segment, a rapid surge of current is generated, manifesting as pulses of microsecond duration that reach peak values on the order of 1 kA. While the instantaneous velocity during the formation of an individual step can approach $10^8\ \text{m s}^{-1}$, the average downward propagation speed of the stepped leader is about $2 \times 10^5\ \text{m s}^{-1}$, meaning the trip between the cloud source and the ground takes approximately 25 ms. Throughout this downward path, the stepped leader branches out into multiple pathways, each of which propagates independently by generating its own series of discrete steps, thereby creating the characteristic ramified structure of the lightning flash.

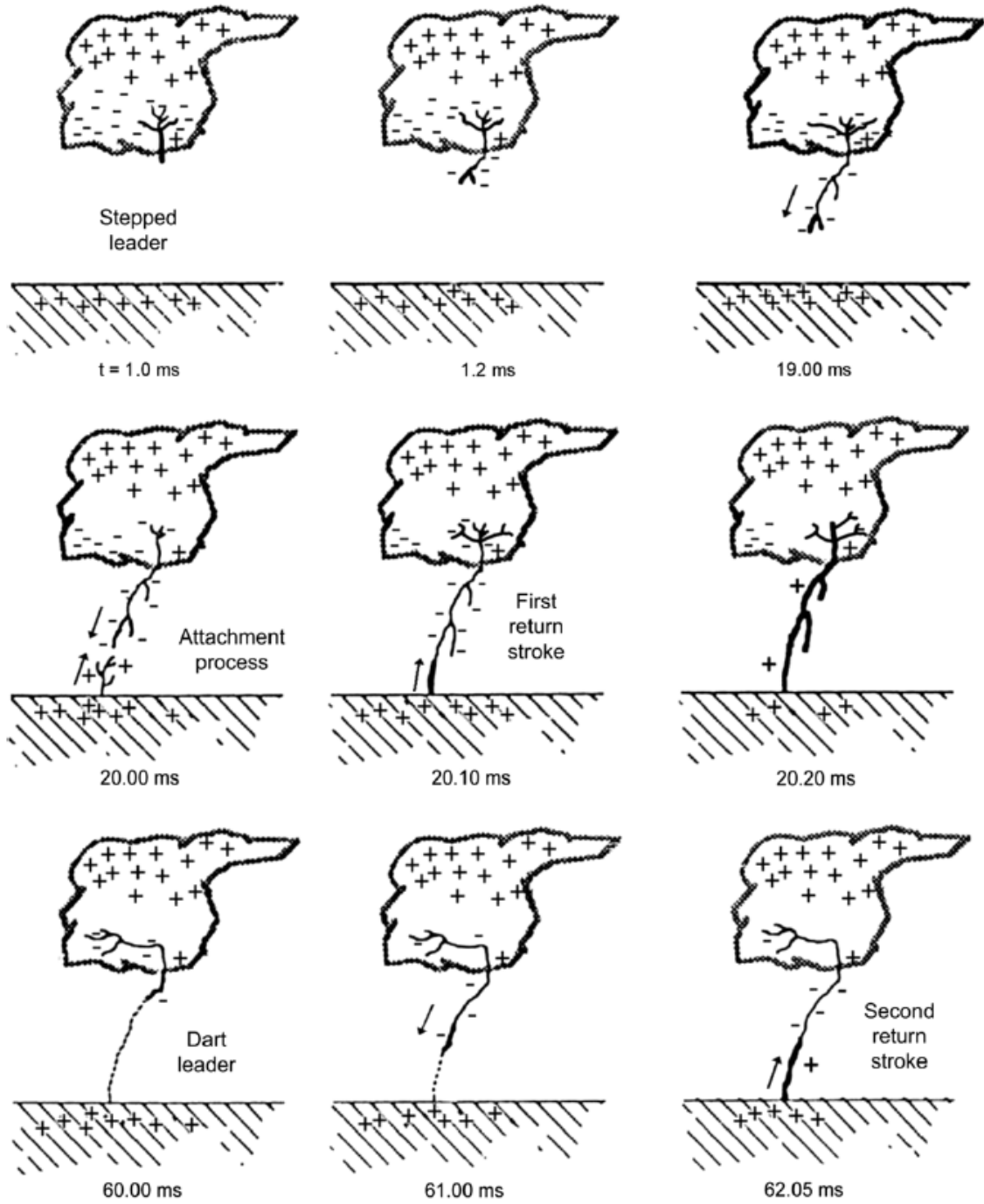


Figure 1.3: Chronological development of a negative cloud-to-ground lightning flash, illustrating the propagation of the stepped leader, the attachment process, the first return stroke, and the subsequent dart leader with respective time scales. *Source:* Reproduced from Dwyer and Uman [11].

The Attachment Process and Upward Connecting Leaders

When the main branch of the descending stepped leader approaches within a few tens of meters of the ground, its negative charge induces a massive positive charge on the Earth's surface directly beneath it. This electrostatic effect is significantly amplified near prominent or elevated structures, such as sharp rocks, antennas and lightning rods. When the localized electric field intensity near the ground becomes large enough, a corona flash composed of weakly luminous filamentary discharges, known as positive corona streamers, is initiated from these grounded structures. As the local electric field continues to intensify, one of these streamers develops into a conductive channel, namely, an upward connecting leader. This upward-going discharge then connects with the branch of the stepped leader, establishing the lightning strike-point and completing the lightning channel between the cloud and the ground, as illustrated in Figure 1.3 at $t = 20.00$ ms.

Return stroke

When the downward-moving stepped leader and the upward connecting leader merge, the negative charge near the bottom of the leader channel moves violently downward into the Earth. This rapid displacement increases the ionization level, driving large currents into the ground and causing the channel base to become intensely luminous. This luminosity and current then propagate continuously up the main channel and down the leader's branches at a speed typically between one-third and one-half the speed of light. This process is the return stroke, illustrated in Figure 1.3 at $t = 20.10$ and 20.20 ms. While the current and luminosity move upward in the main channel, negative charges, the electrons, move downward. Within a few milliseconds of the return stroke initiation, the approximately 5 C of negative charge distributed along the original stepped leader are completely transferred to the ground. Furthermore, once the return stroke reaches the cloud base, an additional current may flow directly from the cloud source to the Earth's surface, contributing further to the total charge transfer of the flash.

The return stroke is responsible for the bright channel of high-temperature air that is commonly seen by the naked eye. Indeed, the maximum return stroke temperature is near 30 000 K, impulsively heating the surrounding air which expands to generate an outward-propagating supersonic shock wave. Upon dissipation, this shock wave transitions into the acoustic wave responsible for the thunder.

From an observer's perspective, the human eye does not visually detect the stepped leader preceding a first return stroke because apparently the visual system cannot resolve the brief time interval—10 ms or less, depending on the channel height—between the formation of the weakly luminous leader and the explosive illumination of the channel by the return stroke. Further, the human eye cannot respond quickly enough to resolve the rapid upward propagation of the return stroke, making it appear as if all points along the channel become bright simultaneously.

Subsequent return stroke and dart leader

After the first stroke current has ceased to flow, the lightning flash may terminate, in which case the discharge is classified as a single-stroke flash. However, about 80% of flashes that lower negative charge in temperate regions contain multiple strokes, typi-

cally ranging from three to five. These strokes are generally separated by an interval of 40 or 50 ms. Strokes following the first, referred to as subsequent strokes, are initiated only if additional negative charge becomes available to the upper portion of the previous stroke channel within approximately 100 ms from the cessation of the preceding current. When this additional charge is available, a continuously propagating leader, known as dart leader, moves down the defunct return stroke channel, in Figure 1.3 at $t = 60.00$ ms. Unlike the initial stepped leader, the dart leader advances smoothly, traveling at an average speed of about 10^7 m s⁻¹ and once again depositing negative charge from the cloud charge region along the channel length (Figure 1.3 at $t = 61.00$ ms). The dart leader thus sets the stage for the second, or any subsequent, return stroke. Because it generally deposits less charge within its channel, perhaps a tenth of that laid down by the stepped leader, subsequent return strokes lower less charge to ground and exhibit smaller peak currents than first strokes. While first stroke currents are typically near 30 kA, subsequent stroke peak currents range between 10 and 15 kA.

The first return stroke in a negative CG flash appears strongly branched downward because it follows the exact path and branches of the preceding stepped leader. In contrast, the dart leader usually follows only the main channel of the previous stroke and hence the subsequent return strokes generally exhibit little to no branching, as shown in Figure 1.3 at $t = 62.05$ ms.

When considering the cumulative effect of the initial stroke, the sequence of subsequent strokes and any potential continuing currents, a typical negative CG lightning flash lowers a total negative charge of approximately 30 C to the Earth, although this value can vary widely depending on the specific event intensity.

Positive Cloud-to-Ground flashes

As previously noted, only 10% of all Cloud-to-Ground discharges are initiated by downward-moving leaders that lower positive charge to the Earth's surface, originating from the upper positive charge region of the convective cloud. From an observational and kinematic standpoint, positive leaders have been studied much less than negative ones. They are inherently less frequent and less luminous, making optical tracking highly challenging. Furthermore, while initial negative leaders consistently propagate via intermittent and discrete stepping, positive leaders exhibit greater variability; their steps are usually less distinct optically, and they can advance either through stepping or in a more continuous path.

A structural characteristic of these discharges is that they usually exhibit a single return stroke, typically followed by a relatively long period of continuing current. Consequently, when a positive leader establishes a connection with the ground, the resulting return stroke displays unique energetic properties. Regarding these current profiles, the median peak current for positive flashes is near 30 kA, similar to that of negative discharges. Positive return strokes can exhibit exceptionally intense peak currents at the ground, exceeding 300 kA, much higher than the peak currents of negative strokes that rarely exceed 100 kA.

As a result of this combination of high peak currents and sustained continuing current, the overall charge transfer in positive flashes can considerably exceed that in negative ones. Although positive flashes are far less common, these extreme electrical proper-

ties present a special hazard that must be taken into account in the design of lightning protection systems.

1.4.3 Transient Luminous Events

TLEs (Transient Luminous Events) are brief emissions of optical light occurring above thundercloud tops in the stratosphere, mesosphere, and lower ionosphere, generated by upward discharges or in response to IC and CG lightning in the rarefied air. There are several types of TLEs: red sprites, TGFs, halos, ELVES, blue starters, blue jets, and gigantic jets. TLEs were observed for the first time in the 1920s, but they were instrumentally recorded only in the early 1990s.

Sprites are red-colored discharges visible to dark adapted eyes and lasting for up to a few tens of ms. They are high altitude discharges associated with lightning occurring in very intense thunderstorms. Halos are brief descending ionization waves producing glows near the base of the ionosphere, with diameters less than 100 km, which sometimes precede or accompany Sprites.

TGFs, standing for Terrestrial Gamma-ray Flashes, originate from thundercloud altitudes, usually in association with IC lightning. They consist of bright millisecond-long pulses of X-rays and gamma-rays. They were detected for the first time by space-based astrophysical observatories, originally dedicated to the study of cosmic gamma-ray bursts (GRBs).

ELVES are rapidly expanding rings of light emission in the lower ionosphere, reaching up to about 300 km across. They are triggered by the electromagnetic pulse from lightning return strokes, that propagates upward as a spherical wave, intersecting the lower ionosphere in a ring that expands outward, heating free electrons and inducing excitation and ionization. The name ELVES is indeed an acronym for Emission of Light and Very Low Frequency perturbations due to Electromagnetic Pulse Sources. ELVES are more common over oceans, presumably because lightning with large peak currents occurs more frequently over water than over land.

Blue starters and blue jets are upward discharges that exit the tops of thunderclouds and terminate in the upper atmosphere, reaching altitudes of approximately 25 and 40 km, respectively. Blue jets propagate upward with speeds of about 10^5 m/s, have lifetimes of about 300 ms and are bright enough to be seen by dark adapted eyes. Gigantic jets are analogous to CG lightning, except that rather than propagating from the thundercloud down to the ground, they propagate upward to the conductive ionosphere at about 90 km [11].

1.4.4 Other Sources of Lightning

Lightning discharges are not uniquely confined to convective thunderstorms. Non-convective phenomena, such as volcanic eruptions and intense dust storms, can also trigger lightning activity. If efficient charge separation mechanisms are active, they can lead to the establishment of an electric field intense enough to initiate electrical breakdowns.

Chapter 2

Instrumentation

This chapter presents a description of the observational systems whose datasets are utilized in this study. An overview of the geographical domain and the spatial distribution of the instrumental network is provided first. Subsequently, a specific section is dedicated to each system: the OTT Parsivel² disdrometer, the LINET lightning detection network and the ARPAE polarimetric weather radars. For each instrument, a brief historical background regarding its technological development is provided, followed by a detailed characterization of its hardware architecture, detection mechanisms and the meteorological quantities retrieved.

2.1 Instrumental Setup and Geographical Domain

The observational analysis conducted in this research focuses on the central-eastern Po Valley (Pianura Padana), a geographical region widely affected by convective phenomena, especially during the warm season. These events are frequently associated with severe hailstorms, downbursts or intense lightning activity. The observational domain specifically centers on the Bologna Metropolitan Area, employing an integrated monitoring framework that combines ground-based in situ instrumentation with atmospheric remote sensing systems alongside a continental lightning detection network. The geographical layout of this monitoring network is depicted in Figure 2.1. The only ground-based sensor is the OTT Parsivel² disdrometer located within the urban core of Bologna at Lat 44.4993° N, Lon 11.3538° E, managed by the Department of Physics and Astronomy (DIFA) of the University of Bologna within the framework of the GID (Gruppo Italiano Disdrometria; www.gid-net.it). It provides direct measurements of the microphysical properties of hydrometeors at the ground level. The available historical archive spans the period from 2018 to August 2025, representing an dataset for evaluating specific convective events over the region. A brief overview of the instrument’s developmental background, along with its features and operating principles, is detailed in Section 2.2.

The atmospheric remote sensing framework consists of the regional weather radar network managed by the Agenzia Regionale per la Prevenzione, l’Ambiente e l’Energia dell’Emilia-Romagna (ARPAE). This network monitors precipitation across the region and provides short-term forecasting (nowcasting), which represent fundamental activities for supporting Civil Protection operations. The infrastructure relies on polarimetric

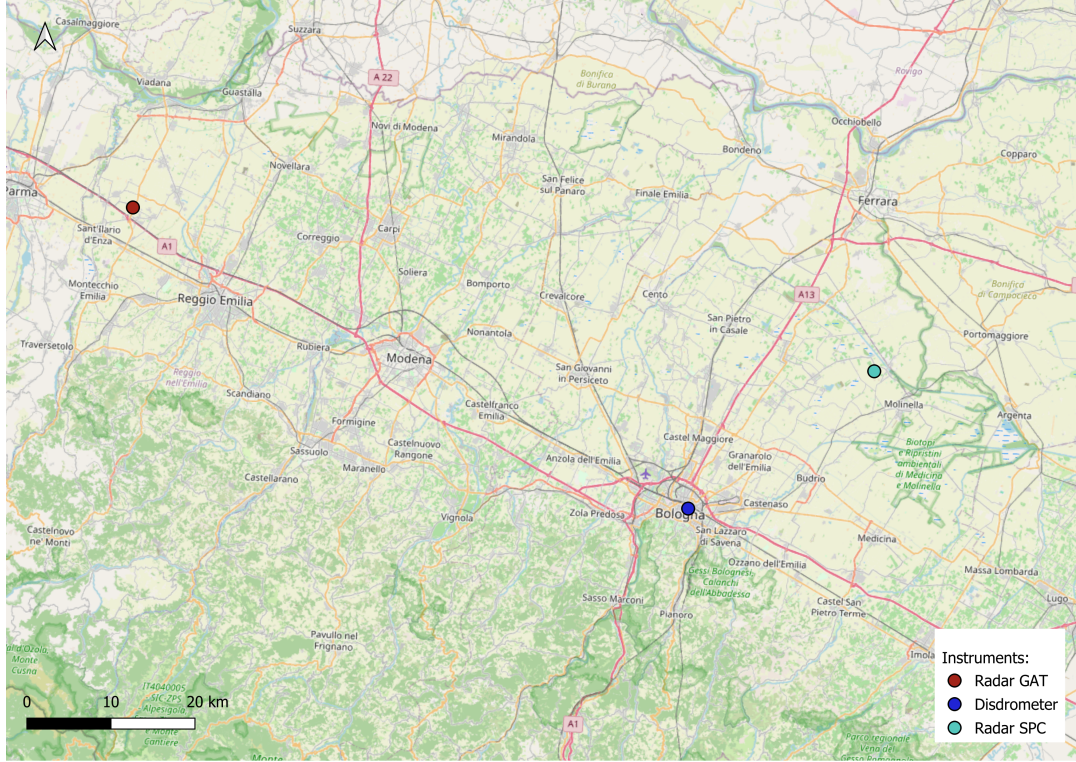


Figure 2.1: Geographical layout of the instrumental setup over the central-eastern Po Valley. Map generated using QGIS.

weather radar systems, positioned at two strategic nodes to monitor the entire regional territory.

- The San Pietro Capofiume (SPC) Weather Radar is sited at the "Giorgio Fea" meteorological base in Molinella (BO) at Lat 44.6547° N, Lon 11.6236° E, with an antenna height of 31.0 m a.s.l. Located approximately 27.48 km away from the urban disdrometer, this system covers the eastern sector of the region.
- The Gattatico (GAT) Weather Radar is positioned in Nocetolo di Gattatico (RE) at Lat 44.7914° N, Lon 10.4992° E, adjacent to the "Fratelli Cervi" Museum, with an antenna height of 60.0 m a.s.l. Located approximately 75.01 km west of the urban disdrometer, this node covers the western sector of the region.

Further details regarding the technical specifications of the systems and their scanning strategies are provided in Section 2.4.

Lightning data are provided by the Lightning detection NETwork (LINET), developed and operated by Nowcast GmbH. At the regional scale, the territory of Emilia-Romagna is monitored through a distribution of sensors, which are dynamically integrated with adjacent stations located in neighboring regions to maximize detection efficiency and location accuracy over the target area. A comprehensive description of the network's technical features and working principles is detailed in Section 2.3.

2.2 The Disdrometer Station

The disdrometer is part of a class of meteorological sensors designed to directly measure both the size distribution and fall velocity of raindrops, in order to characterize the ground-level Drop Size Distribution (DSD) of precipitation, whose formulation and mathematical modeling are detailed in Section 2.2.2.

One of the first systematic approaches to quantify raindrop dimensions was the "flour method", originally introduced by Bentley (1904) and reviewed in detail by Laws and Parsons (1943) [19]. This technique involved exposing a container filled with a thin layer of uncompacted and smooth flour to rainfall for a brief interval to trap drops into dough pellets, which were subsequently oven-dried and weighed to calculate the original drop diameters through empirical mass-size relationships.

Significant technological advancements in the second half of the 20th century marked a transition from manual methods to automated instruments, allowing for systematic and continuous monitoring of precipitation microphysics. Modern automated disdrometers are generally classified into three primary categories based on their operating principles: impact, optical, and video-based systems.

Impact disdrometers infer drop size by converting the mechanical momentum of falling hydrometeors into electrical signals upon physical impact with a sensor head. To overcome the acoustic noise floor, optical disdrometers were introduced, leveraging the attenuation of a laser beam during particle transit to measure both size and terminal fall velocity. For specialized geometric and microphysical analysis, two-dimensional video disdrometers (2DVD) and micro-radar systems can reconstruct detailed particle cross-sections and vertical velocity profiles.

2.2.1 Parsivel² Disdrometer

The Particle Size and Velocity (Parsivel) disdrometer is a laser-optical instrument manufactured formerly by PMTech and recently redesigned by OTT Hydromet into the current model, the OTT Parsivel², an architectural evolution that, as demonstrated by Tokay et al. (2014) [35], enhanced measurement accuracy at both the small and large drop ends. The technical features, structural architecture, and internal signal processing described herein are based on the manufacturer specifications detailed in the instrument documentation and reference manuals [25, 23, 26].

The disdrometer consists of an integrated protective housing enclosing the central processing electronics and supporting two symmetrically sensor heads facing each other, designed to minimize wind obstruction and prevent water dripping into the optical path. One head houses a transmitter unit equipped with a laser diode that emits a visible red beam at a wavelength of 650 nm with a continuous peak output power of 0.2 mW. The laser beam is focused into a flat, horizontal sheet of light that propagates across a horizontal path length of 180 mm to reach a single receiving photodiode housed in the opposite head. This setup defines an active measuring volume 30 mm wide and 1 mm thick, establishing a total sampling surface of 54 cm². The functional mechanism of the instrument is based on the light extinction principle and the shadowing effects caused by hydrometeors passing through the laser band. In the absence of precipitation, the receiver photodiode captures the unattenuated laser sheet, producing a maximum output signal

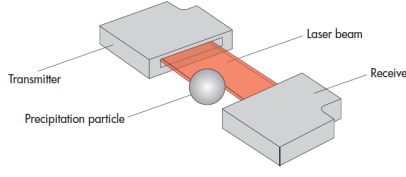


Figure 2.2: Functional Principle of the OTT Parsivel² *Source:* Reproduced from [25].

of 5 V. When a precipitation particle falls through the laser beam, it partially blocks the light strip, causing a transient drop in the voltage output. The degree of dimming, specifically the maximum amplitude of this voltage reduction, is a direct measure of the particle size, from which the volume equivalent diameter of the particle is extracted. The temporal duration of the extinction signal, from the instant the hydrometeor first enters the laser sheet until it completely exits the sampling volume, determines the particle’s terminal fall velocity.

To ensure long-term stability and accuracy, the Parsivel² incorporates an advanced radiometric process that compensates for the physical degradation and aging of the laser diodes, as well as the fluctuations induced by the temperature characteristic curve of the optical sensors.

Data Discretization and Output Framework

The raw data collected during each sampling interval, which features a flexible temporal resolution spanning from 10 seconds to 1 hour and is fixed at 60 s for the dataset analyzed in this study, are processed by a fast digital signal processor. The validated particles are stored in a 32×32 matrix, which consists of 32 uneven size classes with a maximum diameter of 25 mm, and 32 uneven velocity classes covering a interval from 0.2 to 20 m/s. The lowest two size classes, covering diameters below 0.2 mm, are systematically excluded from meteorological evaluations due to an insufficient signal-to-noise ratio close to the electronic noise floor.

From this size-velocity spectrum, the integrated software calculates macroscale parameters, including rain accumulation and intensity, atmospheric visibility, total kinetic energy, and the equivalent radar reflectivity factor, while simultaneously classifying the precipitation type according to the WMO weather codes. All these calculated variables, alongside the raw spectrum, chronological data, and internal instrumental diagnostics, are compiled into a single telegram. This data string can be fully customized by the user to select and output only the specific parameters required for the observational setup.

2.2.2 Drop Size Distribution and its Modeling

The Drop Size Distribution (DSD) is of fundamental importance for the study of cloud and precipitation microphysics. Formally, the DSD is expressed as $N(D)$, and its unit of measure is $m^{-3} mm^{-1}$, which represents the number of drops per cubic meter per millimeter diameter interval, as derived from the raw particle counts of the disdrometer station itself.

The parameterization of the DSD finds applications across various fields of atmospheric sciences, including the numerical modeling of cloud microphysics or mesoscale weather

forecasting. Within this framework, the first analytical parameterization of the DSD was proposed by Marshall and Palmer (1948) [22], by analyzing experimental observations of raindrops collected on dyed filter papers. They observed that the empirical data closely fitted an inverse exponential distribution of the following form:

$$N(D) = N_0 \cdot e^{-\Lambda D} \quad (2.1)$$

where D is the equivalent spherical diameter of the raindrop, expressed in mm . N_0 is the intercept parameter, representing the theoretical number of drops when $D = 0$. Marshall and Palmer (1948) [22] assumed this value to be constant for all rain types, fixing it at $N_0 = 8000 m^{-3} mm^{-1}$. Λ is the slope parameter, expressed in mm^{-1} , which dictates how rapidly the drop concentration decreases as the diameter increases. The parameter Λ is coupled to the rainfall intensity R , expressed in $mm h^{-1}$, measuring the volume of liquid water falling through a unit horizontal area per unit of time, through the empirical relationship:

$$\Lambda = 4.1 \cdot R^{-0.21} \quad (2.2)$$

The Marshall-Palmer (MP) equation describes a decrease in drop concentration, peaking at $D = 0$. While subsequent field experiments have validated the MP distribution as a good approximation for stratiform rain events, Waldvogel (1974) [38] identified systematic shifts in the intercept parameter N_0 , which is not constant but undergoes large variations as a function of the changing precipitation type in convective regimes. To take into consideration these fluctuations, the meteorological community transitioned to the more flexible three-parameter Gamma distribution, a shift definitively formalized by Ulbrich (1983) [37]:

$$N(D) = N_0 \cdot D^\mu \cdot e^{-\Lambda D} \quad (2.3)$$

In this formulation, the DSD is governed by three independent parameters: N_0 , the intercept parameter that scales the overall concentration; μ , a dimensionless shape parameter, whose positive value ($\mu > 0$) generates a downward concavity, modeling the observed depletion of very small drops, while, a negative value ($\mu < 0$) produces an upward concavity. Λ is the slope parameter and is always positive. High values correspond to a steep and narrow distribution characterized by a high concentration of small droplets and an absence of large drops, while low values correspond to a wider and flatter distribution, indicating a spectrum that extends toward larger hydrometeor diameters. This distribution allows the mathematical curve to flex and bend, thereby capturing the natural non-exponential curvature of observed spectra.

The DSD represents the microphysical link to the macroscopic scale; knowing the mathematical distribution $N(D)$ allows for the analytical derivation of all integral parameters, providing the bridge between ground-based and radar observations. Mathematically, any generic integral parameter can be expressed as the n -th moment (M_n) of the DSD, which is defined by the general integration formula:

$$M_n = \int_0^\infty D^n \cdot N(D) dD \quad (2.4)$$

To derive the three parameters of the Gamma distribution (2.3) from empirical data, the Method of Moments (MoM) is widely adopted in literature [8], employing a combination of three statistical moments of the DSD (such as M_2 , M_4 , and M_6 , which will be detailed

in Sec. 3.1).

Different physical quantities correspond to specific mathematical moments (n) of the distribution, as the exponent n represents how the parameter weights the diameter of the hydrometeors. The Rainfall Rate (R), expressed in mm/h , is linked to the 3rd-moment of the DSD, as it depends on the volume of the water drops and it is computed according to:

$$R = 0.6 \cdot \pi \cdot 10^{-3} \int_0^{\infty} v(D) \cdot N(D) \cdot D^3 dD \quad (2.5)$$

where $v(D)$ is the terminal velocity of a hydrometeor of diameter D , expressed in $m s^{-1}$. It is worth noting that if the analytical power-law approximation is substituted for the terminal velocity, e.g., $v(D) \propto D^{0.67}$ as proposed by Atlas and Ulbrich (1977) [3], R becomes directly proportional to the 3.67th moment of the distribution. The Radar Reflectivity Factor (Z), expressed in $mm^6 m^{-3}$, is instead linked to the 6-th moment of the DSD (M_6), according to:

$$Z = \int_0^{\infty} N(D) \cdot D^6 dD \quad (2.6)$$

These two integral parameters are highly correlated via the empirical power-law relation:

$$Z = A \cdot R^b \quad (2.7)$$

Because Z and R depend on different powers of the drop diameter, this relationship is extremely unstable. The pre-exponential coefficient A (ranging from 50 to 500) and the exponent b (ranging from 1.0 to 3.0) fluctuate continuously from one precipitation event to another, or even within the same event, depending on the climate, the season, the hour of the day, the presence of mixed-phase hydrometeors like graupel or hail and measurement uncertainties.

Another integral parameter derived from the DSD is the mass-weighted mean diameter, D_m , expressed in mm, defined as the ratio of the fourth to the third moment of the size distribution:

$$D_m = \frac{M_4}{M_3} = \frac{\int_0^{\infty} N(D) \cdot D^4 dD}{\int_0^{\infty} N(D) \cdot D^3 dD} \quad (2.8)$$

D_m represents the characteristic size of the hydrometeors that contribute most to the total rainwater volume; hence, its shift directly reflects a reorganization of the drop spectrum toward larger or smaller sizes.

2.2.3 Instrumental Limitations and Spatial Representativeness

The evaluation of raindrop size distributions and solid precipitation dynamics requires a careful assessment of the physical and operational limitations inherent to optical disdrometry. Although modern devices provide high-frequency and non-intrusive measurements, their observations are still affected by aerodynamic sampling biases, microphysical sampling errors, coincidence effects, and spatial representativeness constraints.

The primary source of systematic error in disdrometer measurements relies on the interaction between the ambient horizontal wind field and the structure of the instrument itself. Because of its asymmetric design, the disdrometer body acts as an obstacle that significantly distorts the local airflow, generating velocity gradients or turbulences, thereby

altering raindrop fall trajectories. Recently, Chinchella et al. (2025) [10] applied computational fluid dynamics (CFD) simulations to evaluate the wind-induced bias affecting the architecture of the OTT Parsivel² disdrometer. This procedure follows the methodological approach pioneered by Nešpor et al. (2000) [24], who first utilized CFD modeling to demonstrate how structural layout of the 2DVD alters raindrop trajectories near the sensing area.

Through CFD simulations and particle tracking, Chinchella et al. (2025) [10] showed that the disdrometer’s aerodynamic bias depends heavily on wind direction and velocity. When the wind is parallel to the laser beam, the upwind head blocks the airflow, generating turbulence that prevents smaller droplets from entering the sampling area, leading to a severe undercatch effect and thus to an underestimation of their number. Conversely, an oblique wind field (45°) causes irregular particle accelerations that artificially introduce extra raindrops into the sampling area, inducing an overcatch effect and a localized overestimation. The perpendicular configuration leaves the laser beam undisturbed, producing negligible trajectory deflections. Across all configurations, large hydrometeors (exceeding 3 mm) remain virtually immune to aerodynamic effects due to their high mass and terminal velocity.

Another technical constraint inherent to optical disdrometers is the occurrence of coincidence errors. This artifact takes place when multiple particles cross the laser beam simultaneously, or when a trailing hydrometeor enters the sampling volume before the preceding one has fully exited. Since the instrument extracts physical properties based on the total amplitude of voltage drop and the overall blockage duration, a concurrent transition causes the processing unit to misinterpret the combined signal signature as a single large hydrometeor possessing an unphysical fall velocity, distorting the recorded DSD.

Furthermore, a single optical disdrometer provides only a point-like measurement, sampling a millimeter-sized area. To capture the localized evolution of precipitation and mitigate these small-scale instrumental uncertainties, a synchronized network of multiple disdrometers should ideally be deployed during an experimental campaign. However, even high-density ground networks encounter severe limitations regarding regional spatial representativeness, owing to the intense microscale spatial and temporal variability that characterizes rainfall intensity and DSD structures. To transcend these point-source boundaries and capture a truly volumetric perspective of convective cores, these local surface observations should be coupled with atmospheric remote sensing techniques, such as weather radars.

2.3 LINET system

The Lightning detection NETwork (LINET) is a total-lightning detection system developed by Professor Dr. Hans-Dieter Betz and his research group at the Physics and Meteorology Department of the University of Munich (Ludwig-Maximilians-Universität München), whose technical characteristics and operational principles are documented by Betz et al. [4] [5].

The development of the network was driven by the need to overcome the intrinsic limitations of traditional lightning location systems, which struggled both to detect weak atmospheric discharges and to determine emission altitudes.

The final version was validated in 2004 and it exploited modern hardware components and advanced software procedures. In subsequent years, following the technical deployment and managed by nowcast GmbH, the network underwent a massive expansion phase to achieve a quasi-total coverage of the European continent. In Germany, the comprehensive lightning detection network was completed in 2006, setting up an initial group of 30 sensors, followed by an additional 60 sensors set up in 16 neighboring countries, as shown in Figure 2.3. Although the figure depicts the distribution as of May 2008, numerous sensors have been added in the meantime. Today, LINET operates as a dense network over western and central Europe, maintaining an optimized sensor baseline distance of approximately 200 km or less between neighboring stations, which guarantees high-resolution total lightning data even across the Italian peninsula and the Emilia-Romagna region.

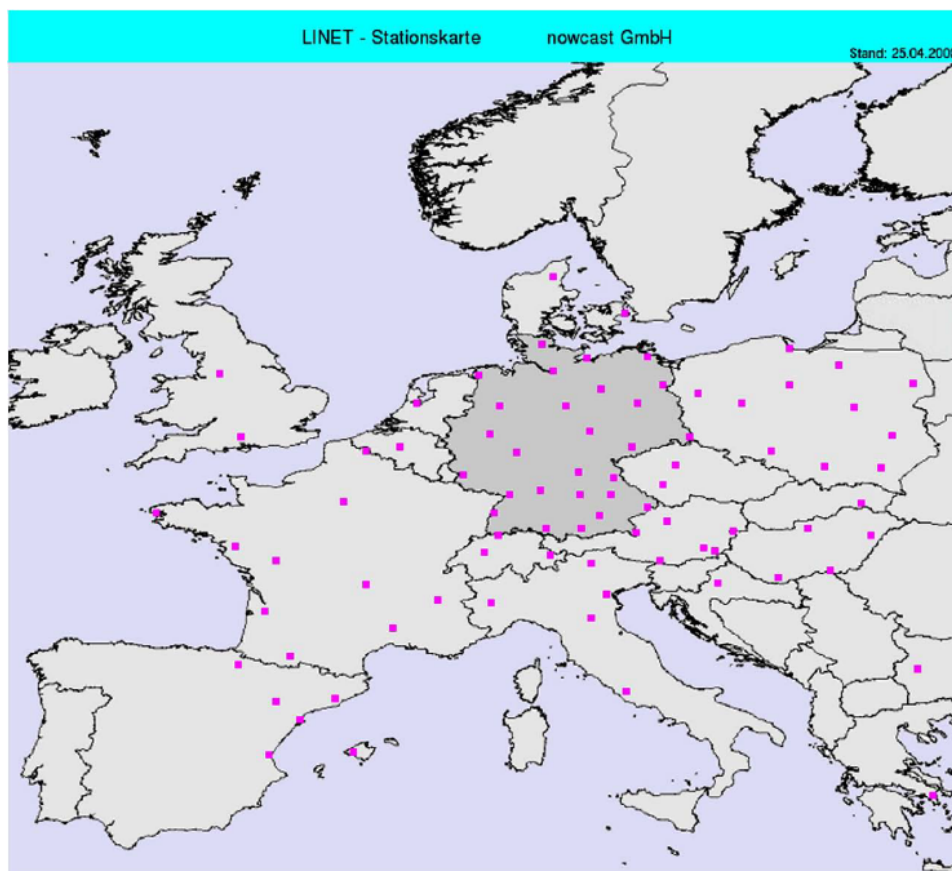


Figure 2.3: Locations of sensor sites as of May 2008. *Source:* Reproduced from Betz et al. [5].

2.3.1 Operating Principles and Detection Efficiency

The operational physics of the LINET system relies on the passive remote sensing of transient electromagnetic pulses radiated during electrical breakdown processes within lightning channels. All network receiving stations feature an identical hardware architecture, as illustrated in Figure 2.4, which displays the sensor antenna and the integrated GPS unit. Each site is equipped with two crossed magnetic loop antennas, designed

to directly measure the horizontal components of the magnetic flux density, $B_x(t)$ and $B_y(t)$, as a function of time from the induced current. While the antenna is capable of capturing frequencies from approximately 1 kHz to more than 1 MHz, the upper limit of the operating range is set to values between 200 kHz and 400 kHz. This spectral window allows the system to operate efficiently within the Very Low Frequency (VLF) and Low Frequency (LF) radio bands. The frequency cut-off serves anti-alias properties and minimizes interference from commercial radio emissions, ensuring that waveform distortions remain within acceptable tolerances for precise pulse analysis. Bearing direction of the incoming electromagnetic wavefront relative to the sensor site is derived from the ratio $B_x(t)/B_y(t)$ and event time tagging is achieved with common GPS receivers of 300 ns accuracy. Sampling is performed at a frequency of 1 MHz in a continuous mode and triggered events are analyzed in a parallel circuit so that no instrumental dead time occurs at the station level.

The waveforms recorded by the synchronized sensor cluster are transmitted to the central processing unit to extract the physical parameters of each individual lightning stroke. The range-normalized peak current (measured in kA) is dynamically inferred from the pulse amplitude.



Figure 2.4: Antenna of a LINET sensor and GPS. *Source:* Reproduced from Petracca [28].

The LINET system introduces a fundamental alternative regarding atmospheric electrical monitoring. Traditional meteorological interpretations believed that the use of Very High Frequency (VHF) techniques to capture intra-cloud discharges was required, assuming that VLF/LF signals from cloud processes are much less intense than those due to CG flashes. VHF radiation is associated with the breakdown of virgin air as part of the channel formation process and leader processes in pre-existing channels. VLF/LF radiation is produced by macroscale lightning processes ranging from long-duration continuing currents, to short-duration transient currents flowing through both cloud and cloud-to-ground channels. However, LINET proves that an optimized VLF/LF system can serve as an alternative for comprehensive total lightning mapping. Comparative field evaluations demonstrate that LINET records a massive volume of IC signals, achieving a

data count that is comparable, or even superior, to the datasets retrieved by specialized VHF arrays.

This capability relies on the fact that internal cloud processes generate VLF/LF pulses with peak currents high enough to propagate and trigger multiple sensors, allowing the system to identify and localize them. The absolute detection efficiency of lightning location networks is governed by a variety of parameters, such as antenna sensitivity, amplitude thresholds, dead-times, signal handling in the receiver, sensor baseline configurations, and central data processing algorithms. In LINET, these hardware and software components have been optimized so that all signals exceeding the localized noise level are processed, provided they hit a specified minimum number of sensors. Consequently, within the optimized layout of the network, where neighboring sensors are below approximately 250 km, currents well below 5 kA are detected and localized.

This exceptional capacity to track low-amplitude events down to small-current values relies on two specific instrumental features. First, the LINET sensor measures the magnetic flux of the lightning signal as a function of time ($B(t)$), rather than the time derivative with subsequent integration, which improves the treatment of weak signals close to the noise floor. Second, all incoming signals are analyzed identically and this is possible because IC-CG discrimination is performed not by means of waveform criteria, but by a specially adopted 3-D algorithm that will be described later.

At the operational level, individual waveforms are processed as soon as a certain amplitude threshold is exceeded, and no pulse is eliminated unless a localized artifact of technical or anthropogenic origin is recognized. While random noise pulses can occasionally bypass this filter and be transmitted to the central processing unit, they rarely produce false "outlier" locations, because random noise lacks spatial coherence across geographically separate sites, thus correlated time-matches among different sensors cannot be found, automatically discarding the artifact.

However, a sampling constraint limits the absolute counting of IC events. While a cloud flash produces abundant internal electrical activity, only the larger signals associated with IC processes featuring peak currents above 1.5 kA have enough energy to be recorded at a sufficient number of sensors required for location. Minor pulses beneath this 1.5 kA threshold are too small to be registered across the minimum sensor cluster. It must be concluded that the true number of IC emissions is significantly larger than the detected fraction within the network dataset, introducing a systematic underestimation of weak cloud processes that must be accounted for during storm cell evaluations.

Optimized network layouts achieve a statistical average 2-D planimetric location error as small as 150 m. This high precision has been experimentally verified by tracking the tight spatial clustering of lightning locations around meteorological and telecommunication towers, confirming the network's capacity to track the development and movement of convective structures.

2.3.2 3-D Lightning Discrimination

Traditional lightning location networks generally execute two-dimensional (2-D) planimetric localization under the assumption that the recorded electromagnetic radiation occurs exclusively near ground level. As mentioned before, IC discharges radiate predominantly in the VHF band; however, they concurrently emit a significant amount of

VLF/LF signals, which tend to overlap with CG amplitude distributions across different current regimes. This overlap makes their discrimination based strictly on amplitude or pulse-shape criteria highly problematic. This introduces a significant systematic error: the lower the current, the more probable the misclassification of positive IC events as positive CG strokes becomes.

To overcome these limitations without requiring additional measurements, LINET exploits a specialized geometrical 3-D version of the Time of Arrival (TOA) method. This technique operates on a simple principle: CG strokes emit VLF/LF radiation dominantly from channel parts close to ground level, whereas IC emissions originate from elevated channel segments inside convective clouds well above the ground. The differences in travel times from the two emission centers are directly exploited within the TOA algorithm, as schematically illustrated in Figure 2.5.

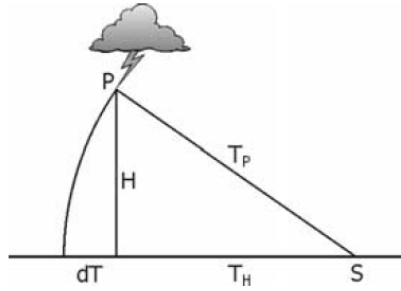


Figure 2.5: Geometric principle of the TOA localization method. P = lightning source, H = emission height, S = closest sensor station, $dT = T_P - T_H$, where dT = residual time delay, T_P = the actual travel time, T_H = theoretical travel time. *Source:* Reproduced from Betz et al. [4].

The locating procedure follows a three-step routine. First, the standard 2-D locating algorithm is applied to determine the initial planimetric ground coordinates of the discharge. Second, for the sensor positioned closest to the lightning event, the algorithm calculates the residual time difference between the actual measured arrival time and the expected value based on ground-wave propagation from the 2-D located point of lightning and the sensor. Third, whenever the emission center is physically elevated, the signal travel path to the nearest station is inherently longer compared to pure surface propagation. Consequently, the arrival time is delayed, allowing the implementation of a dynamic height parameter within the localization routine to solve for the emission elevation. Through this implementation, the locating procedure is carried out directly with four parameters, rather than the three traditional ones (event time and 2-D ground location).

The application of the TOA method depends strictly on sensor availability and geometric constraints. While a minimum of four independent stations allows for the simultaneous determination of both the 2-D coordinates and the emission height, a lesser number of sensors requires TOA values to be supplemented by the station bearing angles. In fact, since the calculated source height is weakly correlated with the planimetric parameters, an inconsistent 2-D dataset cannot be artificially compensated for by introducing a fictitious altitude. Furthermore, the method works efficiently if two conditions are fulfilled: first, the distance between the lightning flash and the closest reporting sensor must not exceed approximately 125 km, requiring sensor baselines of roughly 250 km, otherwise,

the geometric TOA differences become too compressed to be discernible. Second, the location accuracy must be better than the difference of the two relevant travel distances; this condition implies correction of systematic site errors.

While five distinct categories of errors could hamper the 3-D technique, namely statistical timing errors, site and propagation path anomalies, systematic electronic delays, wave-front arrival time identification errors, and sensor saturation effects close to the channel, these factors are actively mitigated through optimized network design and signal processing. Consequently, LINET successfully achieves an effective time accuracy of approximately $0.2\ \mu\text{s}$ (corresponding to a 60m signal travel path) in its central domains, fully satisfying the stringent requirements for reliable IC-CG discrimination.

Through this 3-D algorithm, LINET calculates emission heights that generally lie between 5 km and 10 km altitude. While solutions unphysically close to the ground do not occur, anomalously large heights extending above the cloud-top level are occasionally reported, primarily due to near-field induction effects that lead to an overestimation of the vertical component.

2.4 Weather Radar

The foundations of weather radar technology were laid inadvertently during World War II, when military operators detected the first atmospheric echoes from convective storms using 10-cm radar systems in 1941. Following the war, meteorological studies began to thrive by adapting leftover military radar equipment at pioneering research laboratories. During the 1950s, the field expanded globally, giving birth to "mesometeorology", which represents the link between cloud microphysics and synoptic-scale atmospheric dynamics (Hitschfeld, 1986) [16]. Over the subsequent decades, this technology evolved from the early single-polarization systems into sophisticated dual-polarization networks capable of multi-dimensional quantitative precipitation estimation.

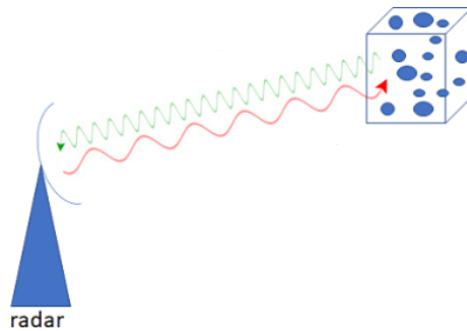


Figure 2.6: Schematic representation of the weather radar operating principle. *Source:* Adapted from ARPAAE [1].

In general terms, active weather radars, where the term "radar" is the acronym for *RAdio Detection And Ranging*, are capable of detecting precipitation in real time, enabling the observation of ongoing atmospheric phenomena within a three-dimensional volume with high spatial and temporal coverage. To achieve this, the radar system utilizes a transmitter that emits focused microwave pulses into the atmosphere and a receiver that

detects the collective energy backscattered by hydrometeors within a discrete sampling volume, whose geometric dimensions expand as a function of distance from the antenna, as schematically illustrated in Figure 2.6. Concurrently, the time delay between pulse transmission and echo reception is exploited to determine the precise distance of the targets, calculated using the known propagation velocity of electromagnetic waves, which equals the speed of light. By analyzing the power, phase, and polarization state of this aggregate returned signal, these systems provide continuous, high-resolution spatial and temporal monitoring of convective cloud structures.

2.4.1 System Hardware Architecture

The regional atmospheric remote sensing infrastructure managed by ARPAAE relies on a dual-node network featuring the two identical ELDES GPM-500C (recently upgraded to ELDES-GPM500CMOD configuration) polarimetric Doppler weather radar systems located at the San Pietro Capofiume (SPC) and Gattatico (GAT) stations introduced in Section 2.1. The system utilizes a high-power Klystron transmitter operating in the C-band with a peak power output of 500 kW at a frequency range of 5.43–5.64 GHz [13]. The choice of C-band for regional network represents an optimal compromise, as it exhibits a contained path attenuation while remaining highly effective at detecting smaller raindrops. The electromagnetic signal is routed through a parabolic antenna with a 5 m diameter, which focuses the energy into a narrow beam with a main lobe width of 0.9° [13]. The antenna performs 360° azimuthal rotations around a vertical axis at sequential elevation angles to execute automated multi-elevation scanning strategies, known as Plan Position Indicator (PPI) sequences. To minimize mechanical wear on the antenna drive system, the radar alternates its mechanical drive between clockwise (CW) and counter-clockwise (CCW) directions. Under severe weather conditions, the scan program consists of 10 discrete elevation steps ranging from 0.5° up to the maximum elevation angle of 25.0° (Figure 2.7), enabling the three-dimensional characterization of convective storm cores and their vertical development. The volumetric acquisition cycle is completed every 5 minutes during convective events (yielding 12 volumes per hour) and decreases to 8 volumes per hour under fair-weather conditions. Depending on the specific scanning mode, the transmitter also adapts its signal properties by varying the pulse width from 0.5 to $3.0 \mu\text{s}$ [13]. This configuration provides a native radial gate resolution of 250 m up to a maximum range of 250 km, ensuring an optimal balance for detecting precipitating structures across the regional network.

2.4.2 The Doppler Effect and Polarimetric Variables

The ARPAAE radar systems exploit the Doppler effect to measure the frequency shift induced by moving hydrometeors, resolving their velocity along the radar line of sight (radial velocity). This Doppler capability is useful for reconstructing vertical profiles of wind speed and direction through Velocity Azimuth Display (VAD) routines.

Beyond these kinematic measurements, the radar network implements advanced dual-polarization technology, transmitting and receiving electromagnetic waves oriented along both the horizontal (H) and vertical (V) polarization planes. This feature allows the system to retrieve information regarding the shape, geometric size, spatial orientation,

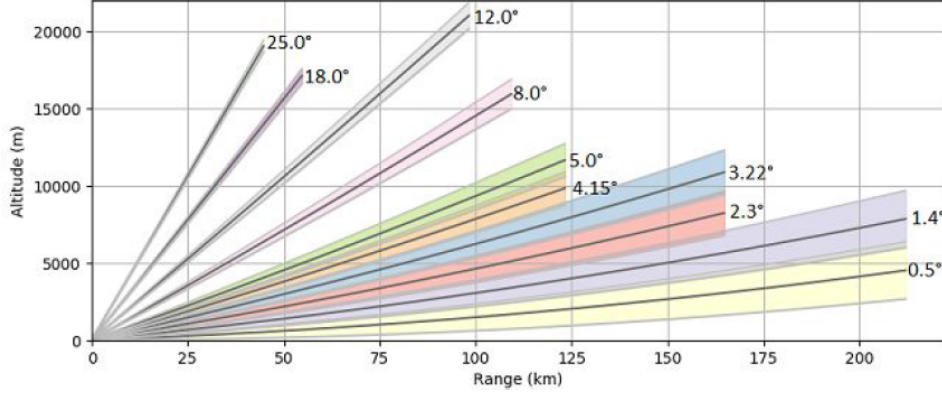


Figure 2.7: Radar operational scanning strategy consisting of ten discrete antenna elevation angles. *Source:* Reproduced from ARPAE [1]

and thermodynamic phase of the intercepted hydrometeors, enabling precipitation type classification. The radar receiver measures the backscattered electromagnetic power, which is then converted into the radar reflectivity factor, Z . As discussed in Section 2.2.2, this factor is directly linked to the instantaneous precipitation intensity, R (measured in mm/h), according to Eq. (2.7). Because the reflectivity factor spans several orders of magnitude, it is conventionally expressed on a logarithmic scale in decibels of reflectivity (dBZ).

By exploiting dual-polarization capabilities, the radar receiver divides the backscattered energy into two orthogonal components. Although the operational network records a wider suite of meteorological products, this study focuses on the variables utilized for identifying solid hydrometeors and characterizing convective structures:

- **Horizontal and Vertical Reflectivity** (Z_H , Z_V), expressed in dBZ, quantify the absolute precipitation intensity based on the backscattering cross-section of the targets along each polarization plane.
- **Differential Reflectivity** (Z_{DR}) allows the discrimination between oblate liquid raindrops ($Z_{DR} > 0$ dB) and tumbling hailstones ($Z_{DR} \approx 0$ dB) and it is defined as follows:

$$Z_{DR} = 10 \log_{10} \left(\frac{Z_H}{Z_V} \right) \quad (2.9)$$

2.4.3 Signal Filtering and BDA Algorithm

Raw radar measurements are intrinsically affected by non-meteorological echoes. Ground clutter (caused by buildings or terrain), sea return, insects, birds, and radiofrequency interference from WLAN networks, alongside anomalous propagation (*anaprop*) under super-refractive conditions can contaminate the reflectivity field. Furthermore, complex terrain, like mountain ranges or hills, can cause partial or total beam blocking, intercepting the microwave pulse and inducing a severe power attenuation that cannot always be fully recovered.

To mitigate these instrumental artifacts, Fornasiero et al. (2006) [13] developed a post-processing technique implemented at ARPAE, called BDA (*Beam blocking, Dynamic*

map, and Anaprop correction) algorithm. The BDA procedure processes the three-dimensional polar volumetric reflectivity data and performs sequential corrections to generate an optimized equivalent reflectivity field, which is then interpolated onto a 2-D Cartesian precipitation map with a 1×1 km resolution.

Instead of assuming standard atmospheric conditions, the BDA algorithm relies on a sequential chain structured into three steps, which simulates the actual radar beam path using refractivity profiles from radiosonde observations. First, a dynamic elevation map selects the lowest clutter-free antenna elevation for each pixel that is simultaneously unaffected by anaprop and significant beam shielding. Second, a vertical continuity test eliminates residual anaprop by comparing the selected reflectivity with higher elevation scans, exploiting the fact that meteorological signals show higher vertical coherence than ground clutter. Finally, a beam blockage correction mathematically compensates for topographic signal attenuation using a geometric optics approach applied to the simulated beam path.

When Fornasiero et al. (2006) evaluated the algorithm on the San Pietro Capofiume radar, the BDA method proved highly efficient compared to traditional static filtering techniques. Ultimately, this procedure yields more robust and realistic 2-D precipitation fields, proving particularly effective in mountain-shielded regions.

Chapter 3

Methodology and Dataset

3.1 Disdrometer Data Processing

Data Extraction, Filtering and Quality Control

The first step of the methodological procedure involves the processing of the dataset provided by the OTT Parsivel² disdrometer. A computational routine is developed to examine the discontinuous historical archive, spanning from December 2018 to August 2025. Specifically, the algorithm sequentially parses the raw telegrams, the data strings generated by the instrument as detailed in Section 2.2.1, to extract the 32×32 spectral matrices. These matrices, which contain the raw hydrometeor counts distributed across 32 diameter and 32 fall velocity classes, are extracted alongside their corresponding chronological and meteorological variables.

A first quality control procedure is implemented to discard invalid or unrepresentative measurements. The raw data are filtered according to three criteria:

- **Non-precipitating intervals:** empty spectral matrices are systematically removed.
- **Instrumental limitations:** spectra exhibiting non-physical raw bin counts, equal to 999, are excluded to avoid the inclusion of instrumental artifacts.
- **Statistically marginal records:** all the spectra containing fewer than 10 total hydrometeors are discarded, as they are considered unrepresentative for subsequent microphysical parameterization.

To ensure that the raw spectral matrices contain actual hydrometeor signatures rather than non-meteorological signals, a velocity-diameter filter is applied. Optical disdrometers are indeed susceptible to measurement errors, such as partially detected drops crossing the edge of the laser beam, splashing effects off the sensor heads, and wind-induced trajectory deviations. To mitigate these anomalies and eliminate spurious echoes, the terminal fall velocity is modeled using the empirical fit introduced by Atlas et al. (1973) in their review paper [2], which approximates the original experimental data gathered by Gunn and Kinzer (1949) [15]:

$$v(D) = 9.65 - 10.3 \cdot e^{-0.6D} \quad (3.1)$$

where $v(D)$ is the terminal velocity expressed in m s^{-1} , and D is the equi-volumetric diameter in mm. Following the quality control criteria established by Tokay et al. (2013) Tokay et al. [36], an acceptance threshold is set to a $\pm 50\%$ deviation from this theoretical curve. Consequently, any particle count recorded at velocity bins falling outside the bounded limits defined by:

$$0.5 \cdot v(D) < v_{\text{measured}} < 1.5 \cdot v(D)$$

is classified as non-meteorological noise and mathematically forced to zero. This should guarantee that only physically consistent measurements are retained, forming the basis for the subsequent derivation of integral variables.

After this initial data refinement, the temporal windows of interest are isolated. The identification of convective episodes is performed by exploiting the present weather codes (SYNOP, METAR and NWS) natively integrated into the disdrometer output telegram. Filtering the dataset according to these indicators allows for the isolation of distinct thunderstorm events for detailed microphysical analysis.

Empirical DSD estimation and Derived Variables

Once the spectral matrices are isolated and filtered, the subsequent step involves the derivation of the Drop Size Distribution (DSD). The valid hydrometeor counts $n_{i,j}$, where the indices i and j denote the diameter and fall velocity classes respectively, are aggregated across all fall velocity bins v_j to calculate $N(D_i)$, which is the number of drops per unit volume and per unit diameter interval expressed in $\text{m}^{-3} \text{mm}^{-1}$. It is computed as follows:

$$N(D_i) = \sum_{j=1}^{32} \frac{n_{i,j}}{A_{\text{eff}} \cdot \Delta t \cdot v_j \cdot \Delta D_i} \quad (3.2)$$

where Δt is the sampling time interval (60 s), ΔD_i is the width of the i -th diameter class and A_{eff} is the effective sampling area of the disdrometer beam. This area is mathematically restricted as a function of the particle diameter to account for partial beam blocking induced by drops crossing the light strip, according to the following relation:

$$A_{\text{eff}} = A \left(1 - \frac{D_i}{2\omega} \right) \quad (3.3)$$

where A represents the geometric sampling area of the laser beam (54 mm^2), and ω is the beam width, which evaluates to 30 mm, as already reported in Sec. 2.2.1.

The three-parameter Gamma distribution, defined by Eq. 2.3, is then employed to model the $N(D_i)$ arrays. The distribution parameters, N_0 , μ and Λ , are estimated through the Method of Moments, which implies that they can be derived from any three selected moments of the DSD. As discussed by Cao and Zhang (2008) [8], the performance of moment-based estimators is challenged by two distinct sources of error: instrumental uncertainties, such as sampling volume limitations, wind-induced noise and measurement thresholds and intrinsic model deviations, which occur when the observed hydrometeor spectra do not strictly conform to the theoretical Gamma distribution. While middle-moment estimators, such as M234, provide optimal results under idealized conditions, they exhibit high sensitivity to these data inconsistencies. On the other hand, estimators

based on higher-order moments are demonstrated to be less sensitive to observational noise and model deviations. Consequently, the M246 estimator, utilizing the second (M_2), fourth (M_4), and sixth (M_6) moments, is implemented in this study. This estimator is strongly recommended by the authors for practical applications on real observational data due to its overall robustness.

Specifically, in the context of convective episodes, the right tail of the DSD is populated by large hydrometeors. Constraining the mathematical fit through the sixth moment (M_6) ensures a reliable representation of these largest particles. The three parameters are thus derived from the selected moments following the mathematical formulation detailed by Cao and Zhang (2008) [8]:

$$\begin{aligned} \eta &= \frac{M_4^2}{M_2 M_6} & \mu &= \frac{(7 - 11\eta) - \sqrt{\eta^2 + 14\eta + 1}}{2(\eta - 1)} \\ \Lambda &= \left[\frac{M_2}{M_4} (\mu + 3)(\mu + 4) \right]^{1/2} & N_0 &= \frac{M_2 \Lambda^{\mu+3}}{\Gamma(\mu + 3)} \end{aligned} \quad (3.4)$$

where η is a dimensionless parameter dependent on the moments ratio, and $\Gamma(x)$ represents the Gamma function.

The empirical distributions enable the derivation of the equivalent radar reflectivity factor Z ($\text{mm}^6 \text{m}^{-3}$), the rain rate R (mm h^{-1}), and the mass-weighted mean diameter D_m (mm), corresponding to the theoretical expressions defined in Section 2.2.2 (Eq. 2.6, 2.5, and 2.8, respectively). These parameters are computed by discretizing those integral formulations into numerical summations over the valid spectral bins and their corresponding terminal velocities.

3.2 LINET Dataset Processing

Characterization and Spatial Constraints

Investigating whether solid precipitation acts as an effective charge-transport vector to the ground requires the characterization of the lightning features in proximity to the surface sensor. For this purpose, data provided by the Lightning NETwork (LINET) are processed and then coupled with the Parsivel² disdrometer measurements. The processing routine imports and handles raw stroke archives delivered in a table format, where each individual entry corresponds to a single detected discharge. The file architecture organizes the physical parameters of each stroke into distinct variables:

- **Temporal Parameters:** The complete date and time information of the discharge, recorded with millisecond precision.
- **Geographical Positioning:** The spatial coordinates, denoting the North latitude and East longitude of the stroke.
- **Emission Altitude:** The vertical location of the discharge expressed in km above mean sea level, which characterizes intra-cloud events, while it is set to zero for cloud-to-ground strokes.

- **Discharge Type Flag:** A numeric identifier utilized to discriminate the category of the discharge, where a value of 1 denotes a CG stroke and a value of 2 denotes an IC stroke.
- **Stroke Intensity (I_p):** The peak current magnitude expressed in kA, whose sign defines the electrical polarity.
- **Quality Indicator:** A statistical parameter reporting the measurement error associated with the spatial reconstruction of the stroke.

The available LINET dataset does not span the entire continental coverage of the network, which extends across the whole of Europe; instead, the raw data were extracted and provided for a restricted sub-domain encompassing the area of interest. This regional domain spans from 44.00° N to 48.00° N in latitude and from 6.00° E to 19.00° E in longitude, covering an area of 445×1000 km, corresponding to 445 000 km². Furthermore, the provided dataset is not temporally continuous, as the data were exclusively extracted in correspondence with the thunderstorm events previously identified and selected from the disdrometer records.

Within this macro-scale domain, the localized electrical activity affecting the sampling volume at ground level is isolated. A specific spatial constraint is thus established by defining a radial tracking zone centered on the disdrometer's location ($\phi_0 = 44.4993^\circ$ N, $\lambda_0 = 11.3538^\circ$ E). The distance between the instrument station and each lightning stroke is computed using the Haversine formulation:

$$d = 2R_E \cdot \arcsin \left(\sqrt{\sin^2 \left(\frac{\Delta\phi}{2} \right) + \cos(\phi_0) \cos(\phi_i) \sin^2 \left(\frac{\Delta\lambda}{2} \right)} \right) \quad (3.5)$$

where R_E represents the mean radius of the Earth (6371.0 km), ϕ_i and λ_i denote the latitude and longitude of the i -th lightning stroke, while $\Delta\phi = \phi_i - \phi_0$ and $\Delta\lambda = \lambda_i - \lambda_0$ are the coordinate differences, expressed in radians, relative to the disdrometer station. Based on the computed distance, a maximum radial threshold is applied to isolate the strokes near the station. This localized domain is required as a specific compromise: it must be sufficiently constrained to capture the electrical activity of the convective core directly affecting the disdrometer's sampling volume, thereby avoiding lightning signals from distant convective cells, while remaining wide enough to ensure a significant sample size for bulk dataset characterization and subsequent probability density modeling.

3.3 Radar Data Processing and Analysis

Radar Data Reading and Acquisition

The radar observations utilized in this study are retrieved from the centralized operational repository of the ARPAE SIMC (*Servizio Idro-Meteo-Clima*) and are integrated into the analysis using two distinct data products. For the mesoscale spatial characterization of the convective events, regional composite datasets are utilized, while three-dimensional volumetric scans are employed for localized vertical profiling.

The radar datasets were acquired exclusively for a targeted subset of the most intense

thunderstorm events. This selection focuses on cases of distinct meteorological and electrical severity, identified at ground level by the simultaneous presence of prolonged solid precipitation, such as hail and graupel signatures in the disdrometer spectra, and intense lightning activity. These specific episodes are designated as primary case studies to undergo deep three-dimensional diagnostic analysis.

The two-dimensional composite maps, provided in the NetCDF (Network Common Data Form) format, are utilized to track and analyze the horizontal evolution of the reflectivity field at the mesoscale. These gridded data combine the observations from both the San Pietro Capofiume and Gattatico radar nodes. This mapping relies on the BDA processing chain managed by ARPAE [13], which integrates specific quality control routines for topographic blockage correction, ground clutter removal, and anomalous propagation filtering, as explained in Section 2.4.3. This procedure collapses the volumetric data into a flat, two-dimensional matrix representing the precipitation field as close as possible to the ground.

To process these datasets and filter out non-meteorological noise, the metadata attributes incorporated within the NetCDF structure are queried. Specifically, the lower detection limit of the radar network is defined by the *detection_minimum* attribute, while the *undetectable* attribute specifies the value assigned to pixels lacking a meteorological return; both are theoretically fixed at -19.69 dBZ. Below this threshold, the backscattered signal is indistinguishable from the instrumental noise floor. In addition, unresolvable or missing data points, caused by total beam blocking or areas outside the radar coverage, are flagged through a dedicated *var_missing* attribute fixed at -20.0 dBZ. To filter out clear-sky background noise and non-meteorological returns, the *valid_min* attribute is retrieved, which represents the lowest reflectivity value actually measured during a specific scan. A thresholding filter is then applied to the reflectivity field, preserving exclusively the grid cells that exceed this minimum value ($Z > \text{valid_min}$), while all pixels below this threshold are discarded.

Construction of Pseudo-RHI and LINET Integration

Weather radar systems sample the surrounding atmosphere using specific scanning strategies tailored to spatial monitoring. One approach is the Plan Position Indicator (PPI) mode, where the antenna maintains a fixed elevation angle while executing a full 360° azimuthal rotation. This sweep generates a conical surface that is subsequently projected as a two-dimensional horizontal map, capturing a planar representation of the atmospheric conditions. The Range-Height Indicator (RHI) mode, on the other hand, locks the antenna at a single azimuthal direction while modulating the elevation angle, providing a vertical cross-section of the atmosphere along that specific sightline.

Since the San Pietro Capofiume (SPC) and Gattatico (GAT) radar operate continuously in PPI mode to guarantee regional coverage, an algorithm is developed to reconstruct vertical cross-sections, designated as Pseudo-RHI, from the raw 3-D volumetric datasets. This tool slices the convective volume along a baseline passing through the disdrometer site, allowing for the inspection of the storm’s internal microphysical structure, such as updraft cores, precipitation shafts, and its correlation with lightning dynamics.

The three-dimensional volumetric scans are provided in the HDF5 (Hierarchical Data Format version 5) format, which is structured hierarchically to handle the multi-elevation radar geometry. Within each file, the raw radar data are organized into distinct groups

corresponding to each specific antenna elevation angle (PPI sweeps). Inside each elevation group, individual datasets store the matrix of raw digital counts for the polarimetric variables, along with their metadata attributes, such as spatial resolution parameters and temporal tracking attributes. Based on this hierarchical structure, the reconstruction of the Pseudo-RHI is carried out through the following phases.

Georeferencing, Ray-Matching, and Altitude Calculation.

The vertical profile plane is defined by establishing a geographic baseline centered on the disdrometer’s location, extending both forward and backward for a certain distance range along a chosen azimuthal angle. This baseline is discretized into a sequence of target points, whose geographic coordinates are determined through geodetic computations on the WGS84 ellipsoid model. For each point along the baseline, the algorithm determines its ground range (d_{radar}) and azimuth ϕ_{radar} relative to the specific radar node. These horizontal coordinates are then matched to the closest radar ray and range gate based on the radar’s angular and spatial resolutions. Once the correct radar bin is identified, its corresponding slant range r is extracted. Finally, the true altitude h above sea level of each radar bin is calculated using the classical 4/3 Earth Radius Model [40]. This equivalent radius formulation accounts for the vertical gradient of the atmospheric refractive index under standard propagation conditions, effectively compensating for bending of the radar beam. The altitude is derived by the following geometric relationship:

$$h = \sqrt{r^2 + (R' + h_0)^2 + 2r(R' + h_0)\sin(\theta)} - R' \quad (3.6)$$

where r is the slant range from the radar, h_0 is the antenna altitude, θ is the spatial elevation angle of the beam, and R' is the modified Earth radius ($R' = \frac{4}{3}R_E$, where R_E is again the mean radius of the Earth).

If these coordinates fall within the valid radar domain, the raw data values are extracted and converted into physical variables (such as horizontal reflectivity Z_H) by applying the metadata calibration parameters:

$$V_{\text{phys}} = \text{count} \times \text{gain} + \text{offset}$$

Concurrently, invalid grid cells corresponding to noise or missing data are filtered out by applying a conditional masking procedure analogous to the quality control algorithm described in the paragraph above. Finally, the extracted scattered radar data points are interpolated onto a regular Cartesian grid using a two-dimensional linear method.

Lightning Activity Synchronization and Spatial Projection.

A temporal and spatial alignment is implemented to integrate the electrical discharges detected by the LINET network. The precise time window of the radar volumetric scan is established by extracting the oldest *startepochs* and newest *endepochs* metadata attributes across all scanned elevation datasets. The lightning dataset is strictly filtered to isolate discharges occurring exclusively within the observational scanning window, ensuring strict simultaneity between the detected convective core structures and the electrical activity.

The synchronized flashes are then mapped onto the vertical profile plane. For each discharge, the algorithm determines its distance and position relative to the disdrometer site, which are then evaluated against the plane’s orientation to derive the perpendicular

offset from the baseline. A filtering threshold is thus applied in order to take into account only discharges falling within a maximum tolerance from the disdrometer-centered baseline for incorporation into the final cross-section.

The accepted discharges are classified by type into IC and CG strokes, and by their electrical polarity. This framework guarantees the simultaneous analysis of the convective storm structure aloft and the corresponding electrical discharge activity both at the surface and within the cloud.

Extraction of the 1D Vertical Profile

To further analyze the vertical structure of convective cells, an algorithm is developed to extract and process volumetric weather radar data, with the aim of reconstructing the Vertical Profile of Reflectivity (VPR) along the atmospheric column above the disdrometer station.

Extracting this vertical profile based only on a single radar pixel above the measurement site relies on the assumption of a perfectly vertical hydrometeor trajectory. However, in real atmospheric conditions, raindrops require considerable fall times to reach the surface and undergo wind drift, which is the horizontal advection by the ambient airflow. This means that precipitation detected at high altitudes by radar above the disdrometer will likely be transported downwind, missing the instrumental sampling area, while the actual hydrometeors sampled at the ground level originate from upper cloud layers located some kilometers upwind. In order to compensate for this offset and lateral displacement, an upscaling approach is implemented by applying a 3×3 pixel spatial window centered over the disdrometer location, expanding the sampling volume into a wider observation column that ensures advected hydrometeors are statistically captured.

Because native radar data are structured within a polar coordinate system defined by azimuth and slant range, the 3×3 window cannot be defined by standard Cartesian squares. The algorithm first identifies the central azimuthal ray and slant-range bin corresponding to the disdrometer site. Then, it includes the immediate adjacent neighbors in both indices, allowing the window to adapt to the polar geometry of each elevation scan.

Calculating the spatial mean within this window introduces a mathematical constraint: because radar reflectivity is expressed on a logarithmic scale, computing a direct arithmetic mean yields physically inaccurate and underestimated results by averaging logarithmic values rather than the actual backscattered energy. Thus, each pixel's logarithmic dBZ value is first converted into the linear radar reflectivity factor (Z , measured in $\text{mm}^6 \text{m}^{-3}$) using:

$$Z = 10^{\frac{\text{dBZ}}{10}}$$

The resulting arithmetic mean computed across the valid subset of these 9 linear Z values is then converted back into the logarithmic scale to construct the vertical profile.

The altitude corresponding to the specific elevation is computed using the standard 4/3 effective Earth Radius Model through Eq. 3.6, as previously detailed for the pseudo-RHI reconstruction. Furthermore, a specific elevation-filtering routine is applied to mitigate ground clutter contamination depending on the radar node. For the SPC radar dataset, the lowest elevation sweep is heavily affected by ground clutter around the disdrometer station; therefore, this first elevation dataset is completely excluded from the profile

reconstruction. On the other hand, because the GAT radar is located further away from the measurement site, its lowest beams traverse the atmospheric column above the disdrometer at a higher altitude, leaving the GAT data free from surface clutter and allowing the inclusion of all available elevation angles.

By iteratively repeating the extraction and averaging procedure for each antenna elevation angle within the radar's volume scan, the vertical profile of reflectivity is finally reconstructed above the disdrometer site.

Chapter 4

Results and Discussion

4.1 Statistical Overview of the Preliminary Dataset

Data Refinement and Filtering Statistics

A statistical evaluation of the raw disdrometer dataset is performed to quantify the overall impact of the quality control process described in Section 3.1. The historical archive, compiled across the entire observation period between December 2018 and September 2025, yields a total of 1,243,754 raw spectra, each representing a 1-minute interval of instrumental sampling.

The overall impact of the sequential data reduction is summarized in Table 4.1. The majority of the records consists of non-precipitating intervals, corresponding to approximately 94% of the total archive, and is therefore systematically discarded. Spectra containing fewer than 10 total drops account for 871 records, while instrumental systematic artifacts, characterized by bin counts equal to 999, are extremely rare, affecting only 6 records across the entire seven-year dataset.

Table 4.1: Global statistics of the sequential disdrometer quality control and filtering procedure.

Spectrum Category	Spectrum Count	Percentage (%)
Total Input Spectra	1,243,754	100.00%
No Precipitation	1,170,883	94.14%
< 10 hydrometeors	871	0.07%
= 999 counts	6	< 0.01%
Completely emptied by v - D	146	0.01%
Total Valid Spectra	71,848	5.78%

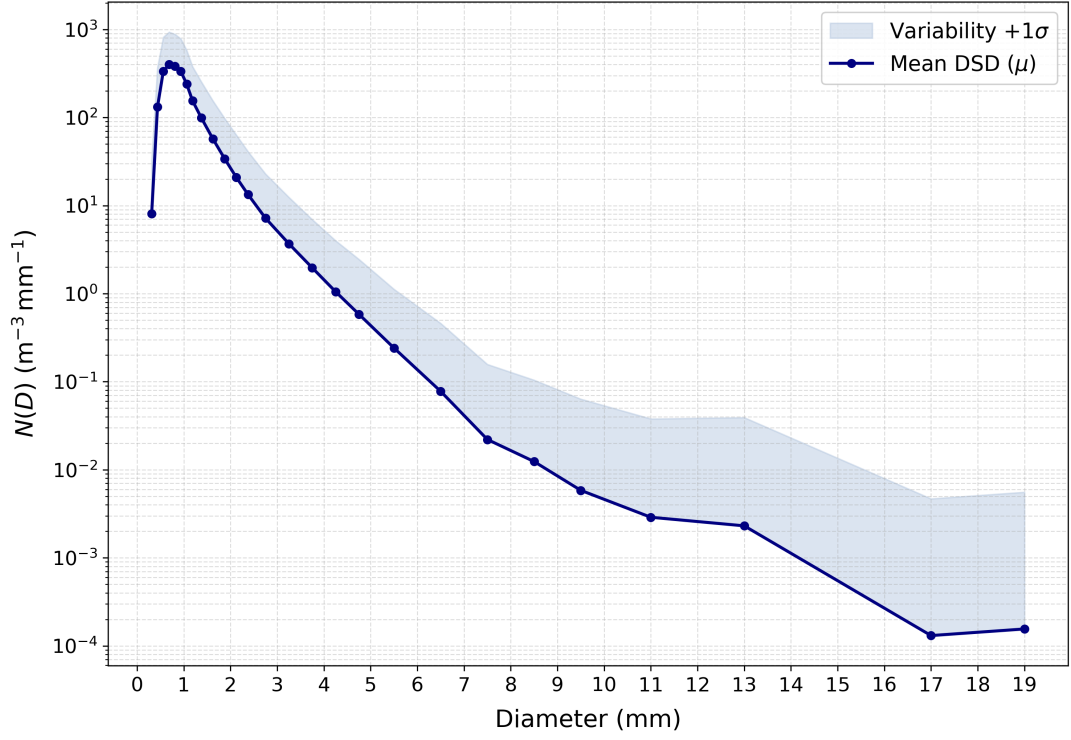
Regarding the application of the velocity-diameter (v - D) filter based on the Gunn and Kinzer formulation (Eq. 3.1), the algorithm proves to be generally effective in removing non-meteorological noise. Out of the active precipitation intervals, only 146 spectra are

completely emptied by the v - D conditioning, while the total number of individual particle counts suppressed is substantial, reaching a cumulative sum of 722,327 filtered drops across the entire dataset.

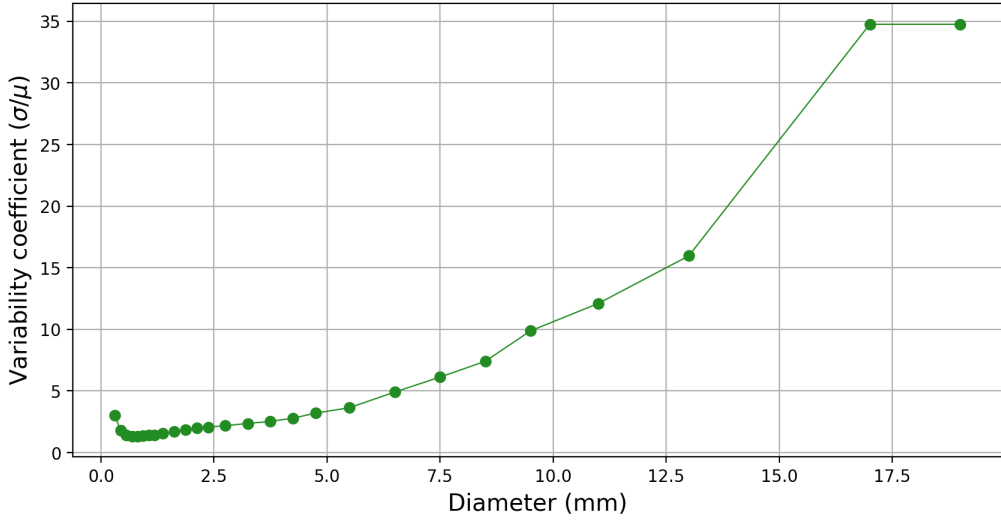
The identification of the convective episodes is achieved by filtering the archive for intervals where the disdrometer recorded solid precipitation. This selection is performed by querying the hydrometeor classification flags, such as SYNOP or METAR codes, which are embedded within the disdrometer’s output telegrams, providing a reliable indicator of hail occurrence. However, since these flags alone do not distinguish between varying degrees of convective severity, a peak rain rate threshold of 45 mm/h is applied as a secondary constraint. This value is established through an exploratory analysis of the long-term archive: by evaluating the global rain rate distribution, 45 mm/h emerged as the optimal boundary to isolate extreme precipitation regimes from the background of moderate convective activity. This filtering strategy yields a final set of 17 major convective events, which are summarized in Table 4.5 in Section 4.1.2. This comprehensive overview detail both microphysical indicators, such as hail duration, and electrical signatures, including lightning counts and peak current intensities.

4.1.1 Microphysical Characterization and DSD extraction

Having isolated the 17 major convective events, an analysis is performed to characterize the spectral evolution and the general microphysical behavior of the selected episodes. The global Drop Size Distribution (DSD) is derived by aggregating all the valid 1-minute spectra from the 17 events, following the procedure detailed in Section 3.1, which enables the conversion of raw instrumental counts into number concentration $N(D)$. For each diameter bin, the mean concentration $\mu_{N(D)}$ and the standard deviation $\sigma_{N(D)}$ are computed to evaluate the spectral dispersion. This statistical approach allows for the quantification of the intrinsic variability of the convective processes, where the standard deviation highlights the range of drop sizes that fluctuate the most throughout the convective life cycle. Specifically, $\sigma_{N(D)}$ is computed across the entire ensemble of 1-minute valid spectra rather than by averaging distributions on an event-by-event basis. This approach is chosen to avoid the artificial smoothing that would result from event-based averaging, which would obscure the rapid evolution inherent in convective storms. By maintaining the 1-minute temporal resolution, the full range of spectral fluctuations is preserved, capturing the transition from developing stages to the mature convective core. Figure 4.1a illustrates the mean DSD, plotted on a logarithmic scale, alongside the spectral variability, defined by the range $\mu + \sigma$. The distribution exhibits the classic convective signature: a rapid increase in concentration for small diameters ($D < 1.0$ mm), peaking at approximately 1.0 mm, followed by an exponential decay for intermediate sizes and a tail which extends toward very large diameters. Regarding the shaded region, it represents only the upper bound of the spectral variability ($\mu + \sigma$). The lower bound ($\mu - \sigma$) is deliberately omitted since the extreme variance at larger diameters ($\sigma \gg \mu$) would yield meaningless negative concentrations that cannot be represented on a logarithmic axis. Observing this upper boundary, it is clear that while the concentration of small droplets remains relatively stable, the spectral dispersion widens significantly for diameters $D > 6.0$ mm. This indicates that the large-particle population is not a permanent feature of the precipitation, but rather a transient component, typical of



(a)



(b)

Figure 4.1: a) Mean DSD aggregated over the 17 convective events (displayed with a logarithmic y-axis), with the shaded region representing the $+1\sigma$ spectral variability; b) Coefficient of Variation (CV) as a function of drop diameter. Y axis is in logarithmic scale.

the intense convective cores. To better isolate this instability, the Coefficient of Variation (CV), defined as σ/μ , is provided and shown in Figure 4.1b. The plot reveals a clear trend: the CV is minimal, close to unity, for smaller drop diameters between 1.0 and 2.0 mm, indicating that the standard deviation is of the same order of magnitude as the mean concentration. This reflects a highly stable background precipitation pro-

cess, where the production of drops is continuous and predictable, ensuring a consistent spectral contribution throughout the convective episodes. In the diameter range 2.0 - 10.0 mm, as the diameter increases the CV grows monotonically, delineating a transition zone characterized by a progressively widening spectral dispersion. For larger diameters ($D > 10.0$ mm), the CV reaches values ranging from 10 to over 30, signaling extreme spectral instability. Within this size regime, precipitation features an impulsive and stochastic behavior: the majority of the event is characterized by the absence of these large hydrometeors, which appear only sporadically in short-lived bursts. This pattern represents the statistical fingerprint of the irregular fallout of large particles, potentially hail or graupel, within the convective core.

It is important to underline that the sharp increase in the CV at the largest diameters is due to a dual effect: it reflects both the actual intermittency of convective storms, where large hydrometeors are highly sporadic, and the statistical limitations of sampling such rare particles with a finite sensing area.

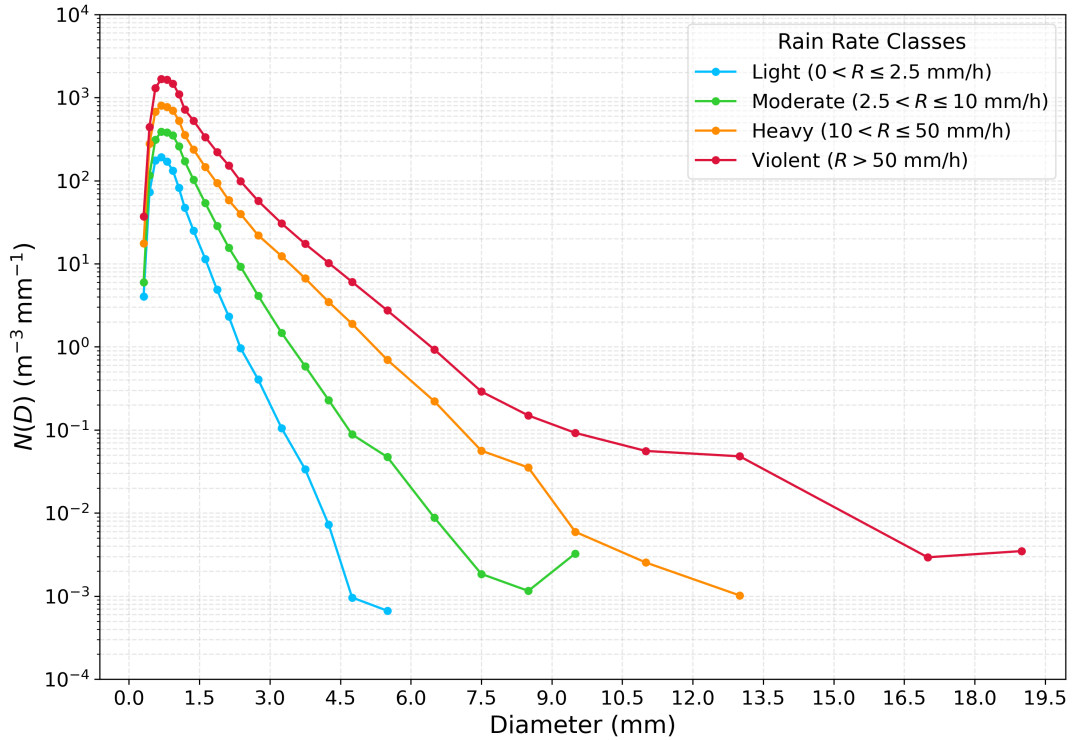


Figure 4.2: Mean DSD categorized by four precipitation intensity classes (Light, Moderate, Heavy, and Violent) defined according to the WMO standards. The vertical axis is displayed in logarithmic scale.

A more detailed assessment of the microphysical properties of the analyzed events is performed by categorizing the valid 1-minute spectra based on the rain rate value, computed as outlined in Section 3.1. In accordance with the World Meteorological Organization (WMO) standards, the following four intensity classes are adopted:

- **Light:** $0 < R \leq 2.5$ mm/h.
- **Moderate:** $2.5 < R \leq 10$ mm/h.

- **Heavy:** $10 < R \leq 50$ mm/h.
- **Violent:** $R > 50$ mm/h.

The resulting mean Drop Size Distributions are presented in Figure 4.2. The plot illustrates a clear evolution of the DSD across the different precipitation regimes. As the rain rate increases, the entire distribution shifts upward, indicating an overall increase in the number concentration $N(D)$ throughout all diameter bins. The slope of the exponential decay at intermediate sizes flattens significantly highlighting a change in the shape of the distribution: as observed transitioning towards the *Heavy* and *Violent* regimes, the spectra are characterized not only by a greater total number of particles, but specifically by a larger contribution from medium-to-large hydrometeors, which progressively broadens the spectrum.

The most notable characteristic is observed in the right tail of the distribution for the *Violent* class: while the *Light* and *Moderate* regimes exhibit a sharp cutoff around 5.0 to 9.0 mm, the DSD for violent precipitation extends further, approaching 20.0 mm. Hydrometeors of this magnitude represent the direct instrumental signature of hail, given that the breakup process prevents liquid water from sustaining such dimensions (Bringi et al, 2003) [6].

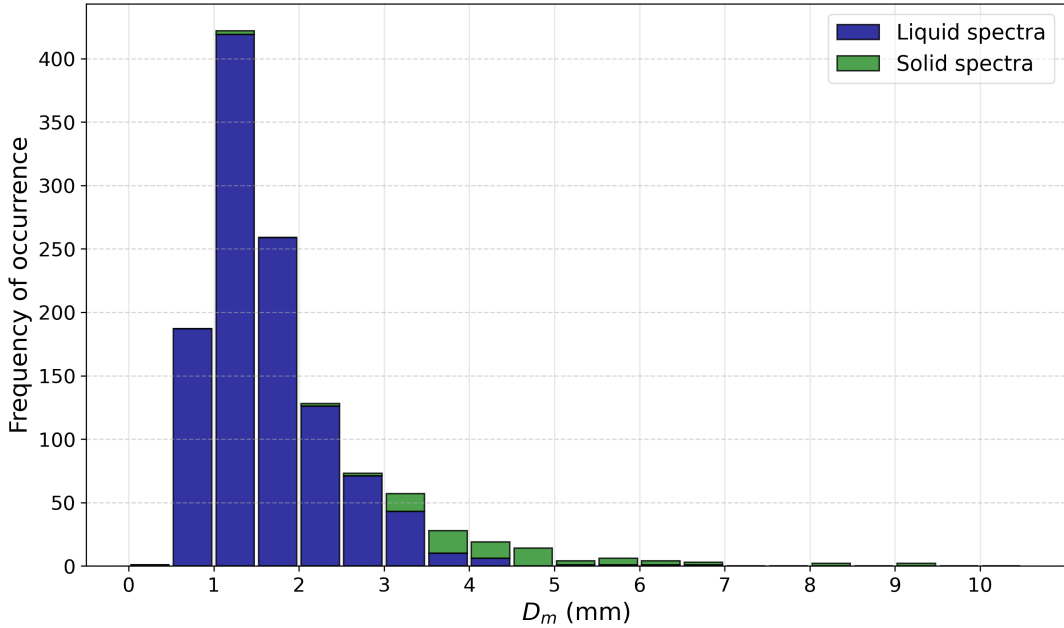


Figure 4.3: Frequency distribution of the D_m , grouped into 0.5 mm bins. The occurrences are partitioned into liquid and solid spectra according to the instrumental SYNOP codes.

To further characterize the differences between the precipitation phases, the frequency distribution of the mass-weighted mean diameter (D_m), computed as described in Section 3.1, is evaluated. Consistent with the previous analysis, this distribution is derived from the valid spectra across the 17 events. Figure 4.3 presents a stacked histogram of the D_m values, computed with a bin width of 0.5 mm and partitioned into liquid and solid spectra based on the instrumental SYNOP codes. The liquid phase heavily dominates the lower end of the size spectrum, exhibiting a pronounced mode between 1.0 and

1.5 mm followed by a rapid decline. In contrast, the contribution of solid precipitation becomes prominent at higher D_m values. Beyond approximately 4.0 mm, the right tail of the distribution is almost entirely composed of solid spectra. This behavior is consistent with the presence of large solid hydrometeors, which shift the mass distribution toward higher mean diameters compared to purely liquid rainfall.

However, a closer inspection of this distribution reveals a subset of nominally liquid spectra exhibiting high D_m values, extending well into the solid-dominated tail. Since the presence of large hydrometeors is the primary driver of such high mean diameters, these anomalies strongly suggest a potential contamination by solid particles within the liquid category.

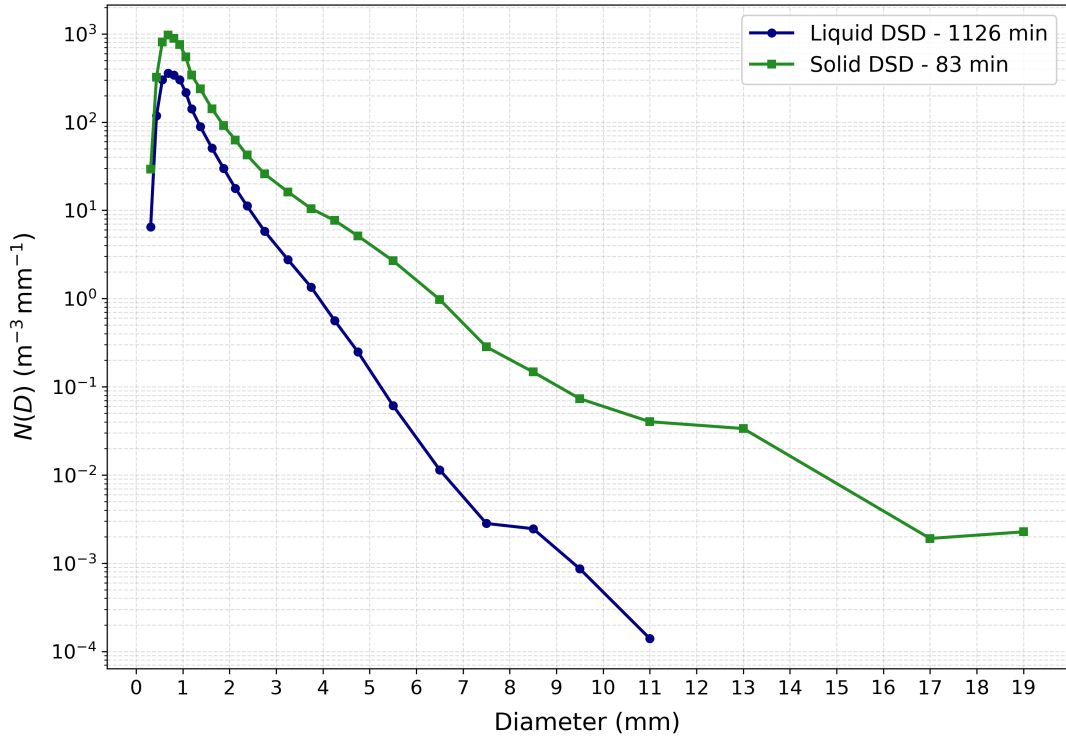


Figure 4.4: Mean DSD computed for the liquid and solid spectra, categorized according to the instrumental SYNOP codes. The vertical axis is displayed in logarithmic scale.

To investigate this hypothesis, the mean DSD for the liquid and solid spectra are compared in Figure 4.4. Both distributions exhibit the characteristic peak at small diameters, ranging between 0.5 and 1.0 mm, followed by a progressive decline, yet their overall shapes present clear differences. The solid DSD (green curve) displays a much broader spectrum with higher concentrations across the mid-to-large diameter range. Its tail extends up to 19.0 mm, which is consistent with the nature of solid hydrometeors: ice particles, such as hail and graupel, can grow to larger sizes by riming and aggregation without suffering from the breakup processes that restrict water drops. In contrast, the liquid DSD (blue curve) exhibits a steeper decay, representing a much faster drop-off in concentration as the diameter increases, reflecting the natural size limits of liquid precipitation.

However, a critical analysis of the extreme right tail of the liquid distribution, where the

blue curve extends up to 11.0 mm, confirms the previously hypothesized contamination. According to the theoretical background outlined in Section 1.1.3, natural raindrops rarely exceed a diameter of 6.0 mm; therefore, the presence of such large hydrometeors within the liquid class implies that solid particles are occasionally misclassified as liquid precipitation by the instrument. In particular, mixed-phase situations involving coexisting liquid and solid precipitation represent an intrinsic challenge for automated sorting algorithms. While these observations expose a limitation in the manufacturer’s proprietary classification system, developing a novel classification scheme is beyond the current scope. Therefore, the original instrumental SYNOP codes are retained for this analysis, acknowledging that a certain degree of solid particle contamination exists within the liquid spectra.

4.1.2 Electrical Activity Characterization

Before evaluating the local electrical response of the storm cloud, the statistical characteristics of the lightning stroke data, provided by LINET network, are initially evaluated on the total aggregated dataset, accounting for all the electrical activity registered across the selected convective events. A preliminary spatial sensitivity analysis is conducted to assess whether the statistical features of the recorded strokes change across different extraction domains. Different radial distances, 50, 30, 25, 10, and 5 km, centered on the disdrometer site, have been evaluated; however, the outcomes are mostly similar across all tested scales, indicating that the statistical behavior of these specific convective events remains uniform from the broader domain down to the local scale. Therefore, to establish a spatially consistent framework for the microphysical comparisons presented later in this study, the lightning strokes within a circular area of 5 km radius have been taken into consideration. This geometric constraint ensures that the analyzed electrical discharges are restricted to the column of air directly encompassing the ground-based disdrometer measurements and the vertical radar pseudo-RHI scans.

To compare these local observations, the dataset is evaluated against the regional baseline established by Petracca et al. (2024) [29], who conducted a 13-year statistical analysis (2010 - 2022) of LINET-detected CG and IC strokes over Italy and the Central Mediterranean area. As detailed in Table 4.2, the stroke partition observed within the 5 km bulk sample, containing a total of 1,983 lightning strokes featuring a 70% CG fraction, represents a stark inversion compared to the Central Mediterranean climatology, where IC pulses constitute the dominant fraction, exceeding 60% of the total network detection. Regarding lightning polarity, negative CG strokes are the most frequent, outnumbering positive CG strokes. This 71% negative fraction is in agreement with the 60% baseline reported in reference literature for the Italian peninsula [29]. The IC population is dominated instead by negative pulses, which account for approximately 74% of the IC total, contrasting with the climatological expectation where positive IC discharges generally prevail (53%).

The Probability Density Functions (PDFs) of the absolute peak currents, categorized by discharge type (Figure 4.5a) and polarity (Figure 4.5b), exhibit unimodal distributions. However, the IC population displays a slight inflection in the decay of the tail beyond 12 kA. The IC population presents a mode between 4 and 5 kA, and is characterized by

Table 4.2: Classification of the total lightning dataset extracted within a 5 km radius. For the total counts, percentages are relative to the grand total of the network, while for positive and negative polarities, percentages refer to the internal fraction of each stroke category.

Lightning Type	Total Count (%)	Positive (+)	Negative (−)
CG Strokes	1,396 (70.4%)	411 (29.4%)	985 (70.6%)
IC Pulses	587 (29.6%)	154 (26.2%)	433 (73.8%)
Total strokes	1,983 (100.0%)	565 (28.5%)	1,418 (71.5%)

a significantly higher peak probability density compared to the CG population, indicating a more constrained range of current intensities for intra-cloud discharges, which are predominantly confined to lower intensity values. The CG distribution exhibits instead a broader shape, peaking at slightly higher values between 5 and 6 kA.

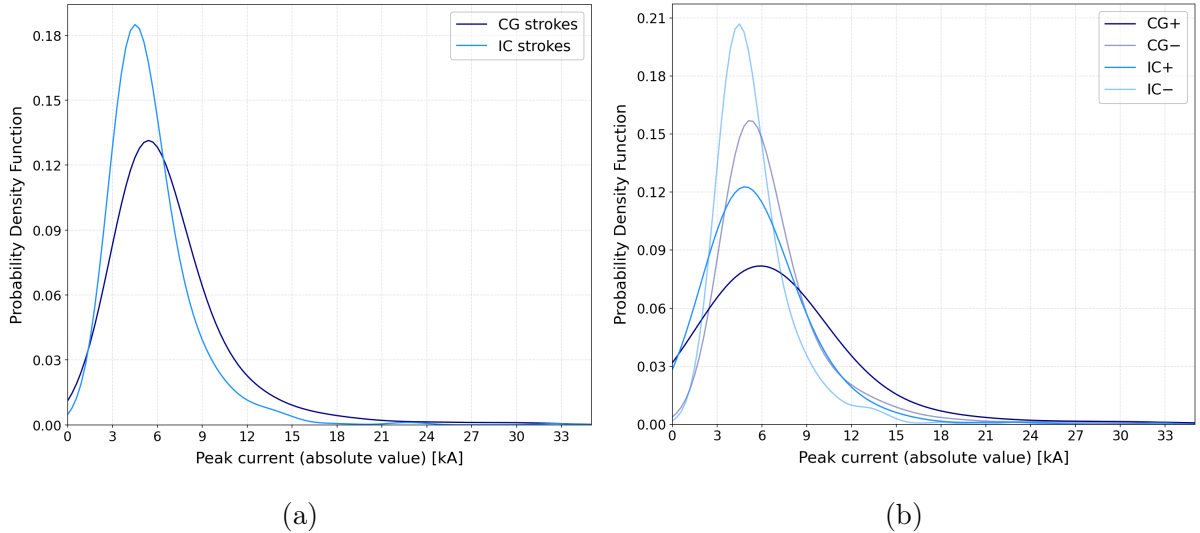


Figure 4.5: a) PDFs (Probability Density Function) of IC and CG strokes current intensities; b) PDFs of IC and CG strokes separated by polarity.

Disaggregating by sub-category (Figure 4.5b), CG− strokes exhibit a narrow distribution with a mode around 5 kA, in contrast, CG+ strokes show a broader distribution, indicating higher variability, with a mode shifted toward higher intensities (~ 6 kA). Regarding intra-cloud activity, IC− strokes display a sharp mode between 4 and 5 kA, and are the primary drivers of the aforementioned inflection observed in the total IC distribution. This feature is likely driven by the aggregated contribution of diverse convective cells within the selected events rather than a distinct secondary discharge process. Finally, IC+ pulses feature a slightly wider distribution peaking at approximately 5 kA.

In general, negative discharges exhibit more pronounced modes at marginally lower peak current intensities compared to their positive counterparts. However, negative discharges also display longer tails in their distribution curves, indicating a greater occurrence of high-intensity events. When compared against the Central Mediterranean climatological baseline [29], the convective events analyzed in this study are characterized by more

energetic discharge processes.

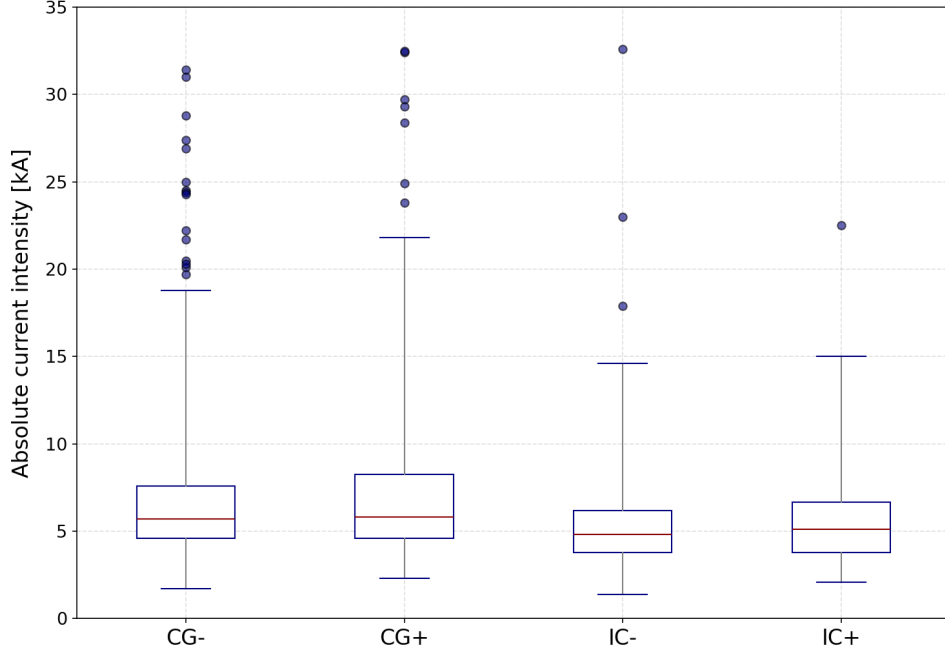


Figure 4.6: Box-plot of IC and CG strokes current intensities separated by polarity. In each box the solid red line indicates the median, and the bottom and top edges of the blue box indicate the 25th and 75th percentiles, respectively.

Table 4.3: Statistical parameters of the absolute peak current intensity distributions for cloud-to-ground strokes and intra-cloud pulses within the 5 km operational radius, separated by polarity.

Statistical Metric [kA]	CG−	CG+	IC−	IC+
Minimum	1.70	2.30	1.40	2.10
First Quartile (25%)	4.60	4.60	3.80	3.80
Median	5.70	5.80	4.80	5.10
Third Quartile (75%)	7.60	8.25	6.20	6.67
Maximum	142.00	142.80	40.60	81.00

To complement the PDF analysis, the absolute peak current intensities of discharge events are evaluated using box plots partitioned by lightning type and polarity, as illustrated in Figure 4.6. The median values remain relatively uniform across all categories, ranging between 4.8 and 5.8 kA. While this is generally within the same order of magnitude as the regional climatology by Petracca et al. (2024) [29], which identifies a broad frequency peak between 3 and 4 kA, the results of this study show median intensities systematically higher across all discharge categories. The interquartile ranges of the positive and negative CG distributions exhibit a larger vertical extension compared to the IC populations, confirming a higher overall variability in ground-discharge energetics. A distinct feature of the present dataset is the distribution of statistical outliers: while

the intra-cloud strokes are generally confined to lower intensities, with absolute maxima reaching 40.6 kA for IC[−] and 81.0 kA for IC⁺, respectively, the CG populations display a longer tail of highly energetic strokes, with absolute peak intensities reaching approximately 142.0 kA for both negative and positive ground strokes (Table 4.3). This behaviour of CG discharges marks a significant departure from the Central Mediterranean baseline, where CG⁺ are predominantly found at lower intensities, while CG[−] display a longer tail in the distribution curve. Furthermore, positive IC pulses feature slightly higher median intensities and a much broader range compared to negative IC pulses, which are more limited in their extreme values.

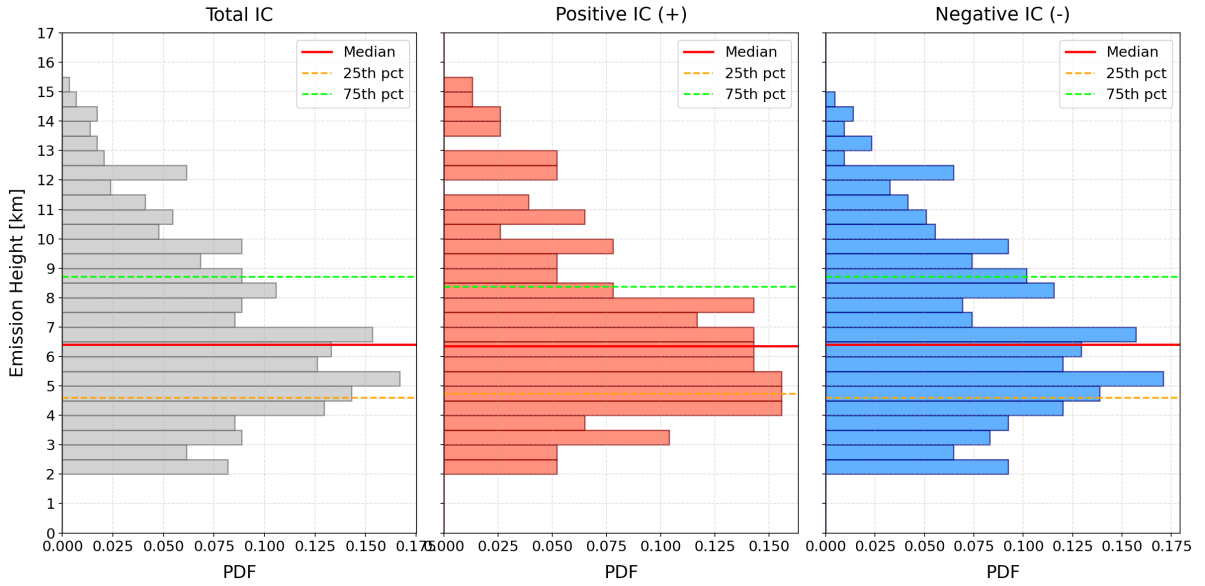


Figure 4.7: Probability density functions of the emission heights for Intra-Cloud pulses. The left panel displays the overall distribution, the center and the right panels show the distributions for positive and negative ICs respectively. Median value, 25th and 75th percentile values are also indicated.

The vertical distribution of the electrical activity is evaluated through the PDFs of IC emission heights within the 5 km radius, computed using a vertical sampling interval of 0.5 km, shown in Figure 4.7. Considering the totality of IC pulses, the recorded emission heights span a range from a minimum of 2.0 km to a maximum of 15.2 km. The dataset displays a median emission height of 6.4 km and features a primary peak concentrated in the 5.0 - 5.5 km altitude layer and a second one in the 6.5 - 7.0 km layer (Figure 4.7, left panel). In comparison to the regional baseline compiled by Petracca et al. (2024) [29], where IC emission heights primarily peak between 6.5 and 8.0 km, the electrical activity in these specific convective events is shifted toward lower atmospheric levels.

The emission height distributions reveal a complex vertical structure, characterized by overlap at mid-levels and a partial stratification near the cloud top (Figure 4.7, center and right panels). Within the central convective body, both positive and negative intra-cloud discharges share an identical median emission altitude of 6.4 km. As detailed in Table 4.4, their interquartile boundaries are not entirely identical: the negative intra-cloud distribution exhibits a slightly broader vertical extension in the middle troposphere, ranging

from 4.6 to 8.7 km, compared to its positive counterpart, spanning from 4.7 to 8.4 km. However, a signature of the stratified charge structure is visible at higher elevations. The overall occurrence of positive ICs is higher at upper-tropospheric altitudes, extending up to 15.2 km and featuring secondary relative maxima at around 12.0-13.0 km. In contrast, negative ICs exhibit a much stronger concentration at middle elevations and are more limited in their upper vertical extent, capped at 14.5 km.

Table 4.4: Summary of the statistical indicators (25th percentile, median, and 75th percentile) for the emission heights of total, positive, and negative Intra-Cloud (IC) pulses.

IC Category	25th Percentile [km]	Median [km]	75th Percentile [km]
Total IC	4.6	6.4	8.7
Positive IC (+)	4.7	6.4	8.4
Negative IC (−)	4.6	6.4	8.7

Table 4.5: Summary of the microphysical features and lightning activity characteristics for the 17 selected convective events.

ID	Date	Time Window	Hail [min]	N _{tot}	IC (-/+)	CG (-/+)	I _{peak}	IC (-/+) [kA]	I _{peak}	CG (-/+) [kA]
Ev. 01	2019-05-28	16:56–17:59	5	38	7/0	24/7	8.2/—	41.8/32.4		
Ev. 02	2019-08-23	15:41–15:57	1	31	11/1	14/5	6.0/5.6	19.7/42.9		
Ev. 03	2019-09-06	07:15–07:53	5	64	10/10	27/17	32.6/22.5	142.0/16.1		
Ev. 04	2020-04-28	11:51–12:55	2	21	8/2	10/1	7.5/13.3	31.0/8.8		
Ev. 05	2020-04-29	12:02–12:11	4	76	27/6	28/15	13.1/5.3	31.4/44.4		
Ev. 06	2020-06-08	16:40–19:29	3	37	5/6	16/10	9.2/4.8	12.2/23.8		
Ev. 07	2020-06-11	13:13–14:42	9	78	23/4	38/13	8.6/6.5	20.3/46.1		
Ev. 08	2021-06-08	13:58–14:42	3	351	124/29	142/56	14.4/11.0	25.0/100.9		
Ev. 09	2024-07-01	16:56–17:29	16	110	7/4	84/15	5.8/12.8	15.3/29.7		
Ev. 10	2024-07-01	19:12–19:39	2	179	14/6	119/40	9.1/7.8	27.4/16.7		
Ev. 11	2024-08-07	17:53–19:42	4	347	80/15	166/86	12.1/7.0	26.9/74.8		
Ev. 12	2024-08-27	12:37–13:10	5	298	40/15	170/73	40.6/81.0	17.5/142.8		
Ev. 13	2024-09-05	17:25–17:52	4	66	8/1	43/14	6.0/4.7	10.6/16.2		
Ev. 14	2024-09-08	17:50–22:46	1	/	/	/	/	/		
Ev. 15	2025-07-08	15:22–16:44	5	60	10/11	24/15	9.7/7.2	24.3/28.4		
Ev. 16	2025-08-20	12:17–13:22	3	12	1/3	3/5	4.5/3.7	14.2/18.2		
Ev. 17	2025-08-21	13:51–15:21	6	215	58/41	77/39	13.3/15.0	22.2/51.0		

4.2 Local-Scale Microphysical and Lightning Activity Evolution

Following the preliminary statistical characterization of the entire 17-event dataset, the focus of the analysis shifts to a localized scale. To investigate the microphysical and electrical coupling during severe convection, a targeted selection of five case studies, has been isolated from the broader sample. The selection of these specific events is driven by a criterion based exclusively on ground-level observational evidence. Specifically, the subset isolates the storms exhibiting the highest combined severity: from a microphysical perspective, events characterized by a prolonged presence of solid precipitation within the disdrometer spectra; and from an electrical perspective, those featuring a high number of lightning strokes, indicative of intense overall lightning activity.

To support these bi-dimensional ground observations with a 3-D diagnostic tool of the internal cloud structure, radar volume data are requested and acquired from ARPAE exclusively for this targeted selection of events.

The selected events are the following:

- **Ev. 08** (June 8th, 2021): a high-frequency electrical event, distinguished by the highest overall lightning count in the entire dataset;
- **Ev. 09 & 10** (July 1st, 2024): a dual-convective sequence characterized by prolonged hailfall and intense lightning activity, providing the opportunity to examine the evolution of two successive storm cells;
- **Ev. 12** (August 27th, 2024): a hailstorm marked by both extended duration of solid precipitation and highly energetic discharge processes;
- **Ev. 17** (August 21st, 2025): a significant convective system featuring a sustained hail phase coupled with widespread lightning occurrence.

4.2.1 Temporal Evolution and Integrated Heatmaps

To examine the ground-level dynamics of these selected events, LINET lightning detections, spatially filtered within a 5 km radius, are temporally synchronized with Parsivel² disdrometer measurements. This synchronization is necessary to account for the different temporal sampling resolutions of the two systems, scaling the lightning data to match the 1-minute integration time of the disdrometer. The temporal evolution of the hydrometeor DSD is represented through integrated heatmaps, where drop number concentrations are expressed on a logarithmic scale, onto which integral variables, computed as detailed in Section 3.1, are overlaid to illustrate the coupling between microphysics and electrical activity. Specifically, two time-series representations are analyzed for each event: the first shows the temporal evolution of the DSD together with the precipitation rate (RR) and the absolute peak current of individual lightning strokes, while the second aligns the mass-weighted mean diameter (D_m) with the lightning stroke frequency. In both visualizations, lightning activity is categorized by discharge type and polarity. Furthermore, the specific time intervals during which solid precipitation is detected by the sensor are highlighted with shaded vertical bands, allowing for a clear visual identification of the hail stages.

Ev. 08 - June 8th, 2021.

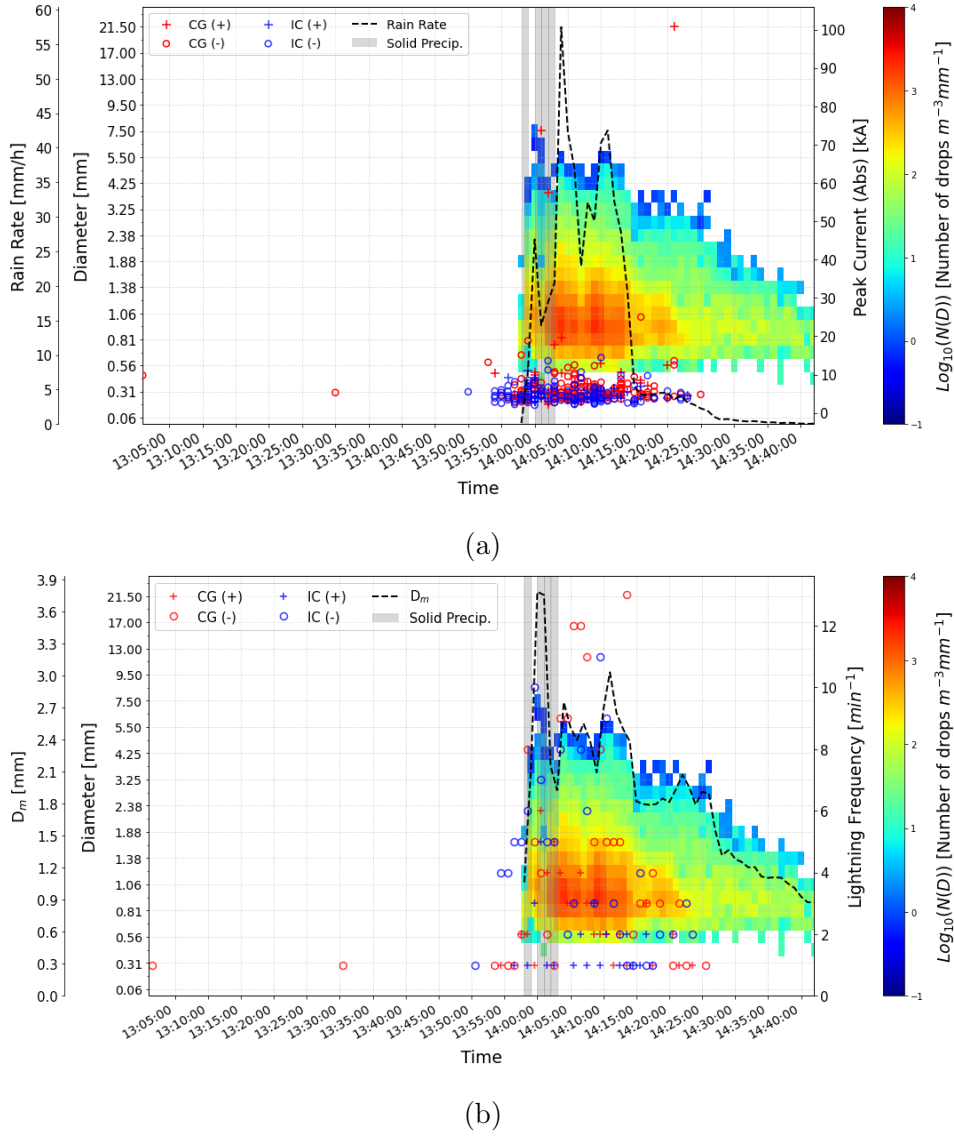


Figure 4.8: Temporal evolution of the logarithmic drop size distribution for Ev. 8. Panel (a) shows the RR profile overlaid with the absolute peak current of lightning strokes. Panel (b) displays the D_m overlaid with the lightning stroke frequency (strokes min^{-1}). In both panels, lightning activity within a 5 km radius is categorized by discharge type and polarity.

The event under analysis lasts around 40 minutes, whereas the lightning activity spanned 25 minutes, with an isolated stroke detected approximately 50 minutes prior to the onset of the event detected by the disdrometer. Lightning activity is concentrated during the mature phase of the storm. At the storm's onset, the drop size distribution is dominated by large raindrops and hail, coupled with a lack of smaller hydrometeors (Figure 4.8). This configuration drives the mass-weighted mean diameter to its absolute maximum (Figure 4.8b), a phase that directly coincides with the occurrence of the highly energetic cloud-to-ground strokes depicted in Figure 4.8a. Minutes later as solid precipitation ceases, a sharp increase in small-drop concentrations triggers the precipitation rate peak

(Figure 4.8a), simultaneously lowering the overall D_m value due to the heavy statistical weight these numerous drops carry in the computation. During this transition, the lightning stroke frequency reaches instead its absolute maximum, coinciding with the RR peak, with a surge in the discharge rate to approximately $13 \text{ strokes min}^{-1}$ (Figure 4.8b). This peak phase is dominated by negative cloud-to-ground strokes and intra-cloud activity. The observed temporal lag between the hailfall and the peak electrical activity is in contrast with the expectations. Indeed, assuming solid precipitation acts as the mechanical vector for downward charge transport, the fallout of hail observed at the storm's onset should theoretically deplete the charge reservoir within the thundercloud, leading to a drop in lightning frequency. Instead, the maximum discharge rate is achieved after the solid precipitation detection. Following this intense phase, the cessation of lightning activity coincides with a significant decrease in both drop diameters and concentrations. However, relying exclusively on single-point ground measurements prevents a comprehensive understanding of these dynamics due to the limited spatial representativeness of the disdrometer. Because the instrument only samples the air column above the station, the data cannot clarify whether the delayed electrical peak is driven by continuous charge separation in adjacent regions or by nearby convective cores; similarly, it is impossible to determine if the final decay signature reflects the actual dissipation of the local updraft or the advection of the active cell outside the 5 km domain. To resolve these ambiguities, ARPAE volumetric radar data will be utilized in the subsequent analysis to trace the broader three-dimensional evolution of the storm.

The greatest raindrop dimensions, up to the exceptional 7.5 mm, are recorded during the initial stage of the precipitation event, followed by a gradual decrease in raindrop diameter in the subsequent minutes along with the decrease in precipitation intensity, thus defining the dissipation of the precipitation. A similar trend is observed for drop concentrations: maximum values for small drops in the 0.6-1.4 mm range are confined to the first minutes of the event, after which they also experience a progressive decline.

During the final stage of the storm, which lasts approximately 15 minutes, no peaks are observed in the RR and D_m values, both of which display instead a steady decay, characteristic of the stratiform-like tail of the convective cell.

Ev. 9 - July 1st, 2024.

The disdrometer records a total duration of approximately 30 minutes for this convective episode, with lightning strokes detected throughout the entire period of time. The onset of the precipitation event is characterized by relatively large hydrometeors in the diameter range of 1.1 – 5.5 mm, whereas small raindrops are completely absent. In the subsequent minutes, the detected diameter range broadens significantly, with hydrometeors reaching a maximum dimension of up to 17.0 mm during the height of the convective phase. This stage coincides with the maximum RR peak, the highest recorded among all the analyzed events, reaching 100 mm h^{-1} (Figure 4.9a), occurring simultaneously with the D_m peak, which attains a maximum value of 9.0 mm (Figure 4.9b). Specifically, these largest hydrometeors are hailstones, as their size heavily exceeds the physical stability limit for raindrops, preventing them from existing in liquid form. This episode features the most intense hailfall of the entire dataset, characterized by 16 minutes of solid precipitation detected by the disdrometer (Figure 4.9). Additionally, extremely high drop concentrations characterize this stage for the smallest drops, in the diameter

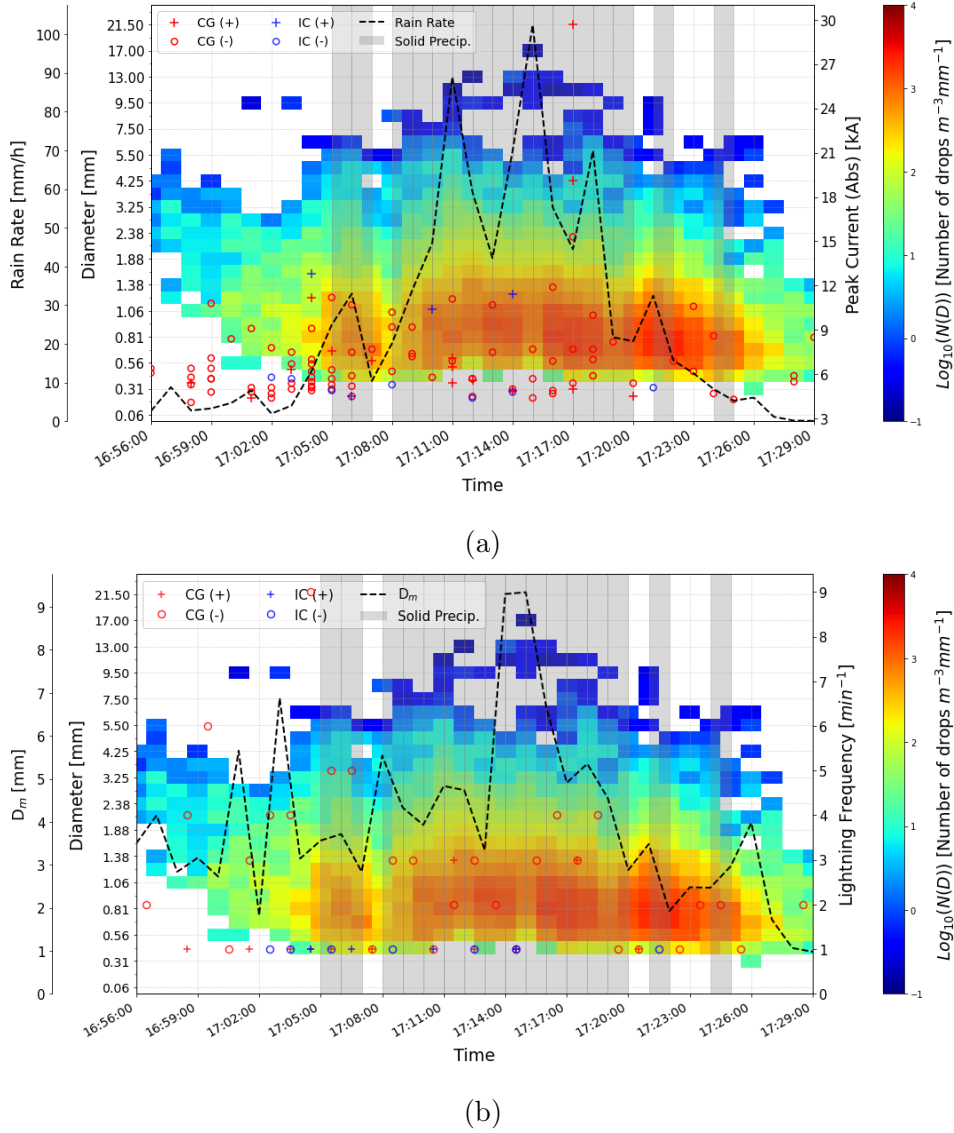


Figure 4.9: Temporal evolution of the logarithmic drop size distribution for Ev. 9. Panel (a) shows the RR profile overlaid with the absolute peak current of lightning strokes. Panel (b) displays the D_m overlaid with the lightning stroke frequency (strokes min^{-1}). In both panels, lightning activity within a 5 km radius is categorized by discharge type and polarity.

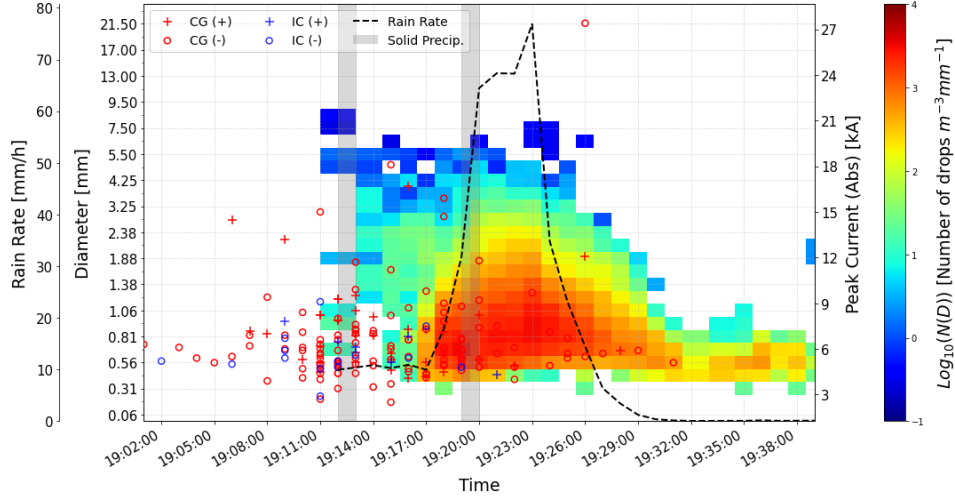
range of approximately 0.5 – 1.4 mm, while the largest hydrometeors naturally exhibit lower concentrations.

The mature convective phase is much more prolonged compared to the weakening phase, which is particularly brief. This results in a rapid depletion of the storm, with the integral quantities dropping to their lowest values. Once again, due to the point-measurement nature of the disdrometer, whether this cessation is due to a dissipation of the updraft or a swift passage of the storm core over the sensor cannot be determined.

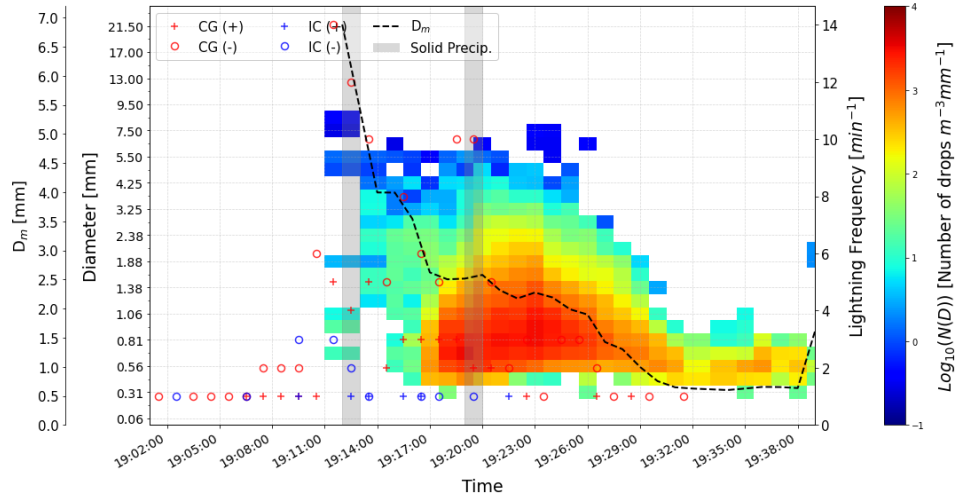
Regarding the electrical activity, the most intense lightning strokes, featuring an absolute maximum peak current of 30 kA, occur during the core convective phase of the storm, coinciding with the hailfall detection. When the solid precipitation ceases, lightning activity decreases significantly in both intensity and frequency (Figure 4.9). This behav-

ior further corroborates the hypothesis that precipitating hail facilitates the downward transfer of electrical charge, mitigating the cloud's electrification.

Ev. 10 - July 1st, 2024, 2nd event.



(a)



(b)

Figure 4.10: Temporal evolution of the logarithmic drop size distribution for Ev. 10. Panel (a) shows the RR profile overlaid with the absolute peak current of lightning strokes. Panel (b) displays the D_m overlaid with the lightning stroke frequency (strokes min^{-1}). In both panels, lightning activity within a 5 km radius is categorized by discharge type and polarity.

This second event represents a subsequent pulse likely generated by the same organized convective system responsible for the previous episode, occurring about 100 minutes after the cessation of the first event (which ended at 17:29 UTC, while this pulse commenced at 19:12 UTC).

As captured by the Parsivel² disdrometer, the evolution of this event exhibits a distinct

mature phase lasting approximately 20 minutes, followed by a rapid decaying phase of nearly 10 minutes. Lightning activity does not occur throughout the entire event but is confined to a time window starting even before the onset itself and concluding with the cessation of the convective phase. The peak current of lightning strokes exhibits a peculiar behavior relative to the hail stages. Rather than during the core of the solid precipitation pulses, an energetic activity, featuring cloud-to-ground strokes with peak currents up to approximately 18 kA, is recorded within the temporal gap separating the two distinct hail bands (Figure 4.10a). Conversely, during the brief hail falls themselves, the stroke currents remain below 10 kA. The absolute maximum peak current of the event, consisting of a negative CG stroke reaching 27 kA, is detected some minutes after the hail stage has completely concluded. The lightning frequency peaks exactly during the minutes when solid precipitation is detected, reaching its maximum flash rate of 14 strokes min^{-1} , as illustrated in Figure 4.10b. Immediately following the second minute of hailfall, a sharp decline in electrical activity is observed, with frequencies decreasing as the storm progresses. This trend aligns with the expectation that the fallout of hail carries charges downward, effectively depolarizing the cloud.

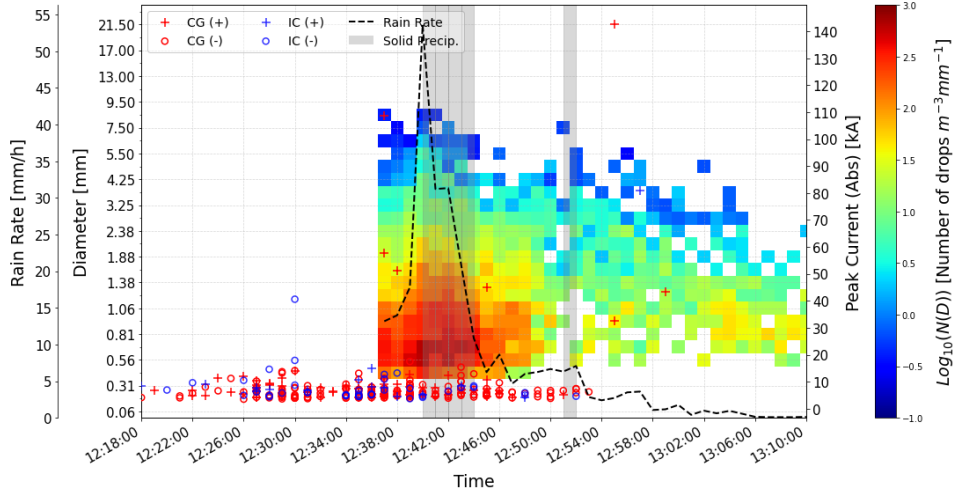
The D_m peak, capped at 6.5 mm, lies where the largest drops, hailstones and lower small-drop concentrations are simultaneously registered during the first minute of the ground-based sensor's record, as shown in Figure 4.10b. The D_m value then starts to decrease as the spectrum becomes heavily dominated by the smaller drops; this decline in D_m mirrors a sharp increase in small-drop concentrations, in the diameter range 0.5 – 2.0 mm, which drives the RR up to its maximum peak of approximately 76 mm/h, after the brief hail stage has concluded (Figure 4.10).

In the weakening phase of the storm, lower drop concentrations and much smaller raindrops are found in the diameter range 0.3 – 1.0 mm. No significant peaks are observed in the RR and D_m values, both of which exhibit relatively flat profiles. Nevertheless, an anomalous signal indicating larger drops is recorded during the final minutes of the precipitation event, which, considering the low concentrations across the entire spectrum, suggests a possible instrumental error.

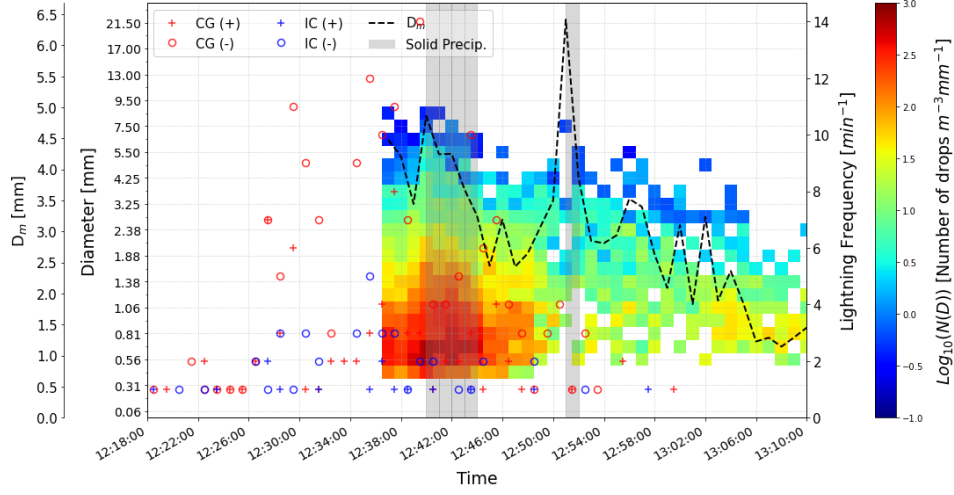
Ev. 12 - August 27th, 2024.

While the precipitation event, occurred on 27th August, 2024 and captured by the disdrometer, lasts approximately 30 minutes, the associated lightning activity spans nearly 40 minutes, with the first strokes detected 20 minutes before any hydrometeors reach the ground (Figure 4.11a). The typical features of a storm onset, specifically the initial arrival of solely large raindrops and the lack of smaller ones, are not observed here. Instead, the record begins abruptly with high hydrometeor concentrations and large dimensions. The maximum raindrop diameters, reaching up to 9 mm, are recorded in the opening minutes of the precipitation event, corresponding to the mature convective phase of the storm. This is attributed to the fact that the disdrometer was activated after the storm had already begun, which also explains why significant lightning activity occurs well before the start of the precipitation recording.

As the event progresses, a gradual decrease in raindrop diameters is observed, marking the transition toward the apparent local dissipation phase of the storm, subjected to the same spatial ambiguities discussed for the previous events. A similar trend characterizes drop concentrations: the maximum values are observed during the first minutes of the



(a)



(b)

Figure 4.11: Temporal evolution of the logarithmic drop size distribution for Ev. 12. Panel (a) shows the RR profile overlaid with the absolute peak current of lightning strokes. Panel (b) displays the D_m overlaid with the lightning stroke frequency (strokes min^{-1}). In both panels, lightning activity within a 5 km radius is categorized by discharge type and polarity.

record for the smallest drops in the diameter range $0.4 - 1.5 \text{ mm}$, while a significant reduction is registered as the storm weakens.

Regarding the electrical activity, it persists for approximately 15 minutes after the precipitation record begins. The stroke frequency reaches its peak of about $12 \text{ strokes min}^{-1}$ before the beginning of the storm (Figure 4.11b), where a cluster of highly energetic strokes is registered, preceding the first detection of solid precipitation by a few minutes. Notably, this event exhibits the highest peak current among the analyzed cases, with a CG stroke exceeding 140 kA detected during the final phase of the storm, as shown in Figure 4.11a. As the hailfall ceases, the lightning frequency begins a slight decline, ultimately fading as the microphysical parameters significantly decay during the final

minutes of the dissipation phase.

Furthermore, RR and D_m peaks are again temporally offset. The RR peak occurs during the convective phase, characterized by the simultaneous presence of hail, large raindrops, and an intense concentration of small drops, whereas the D_m peak lies at the transition toward the weakening phase, where large raindrops still persist but very low small-drop concentrations are recorded, driving the D_m value upward before the final decay.

Ev. 17 - August 21st, 2025.

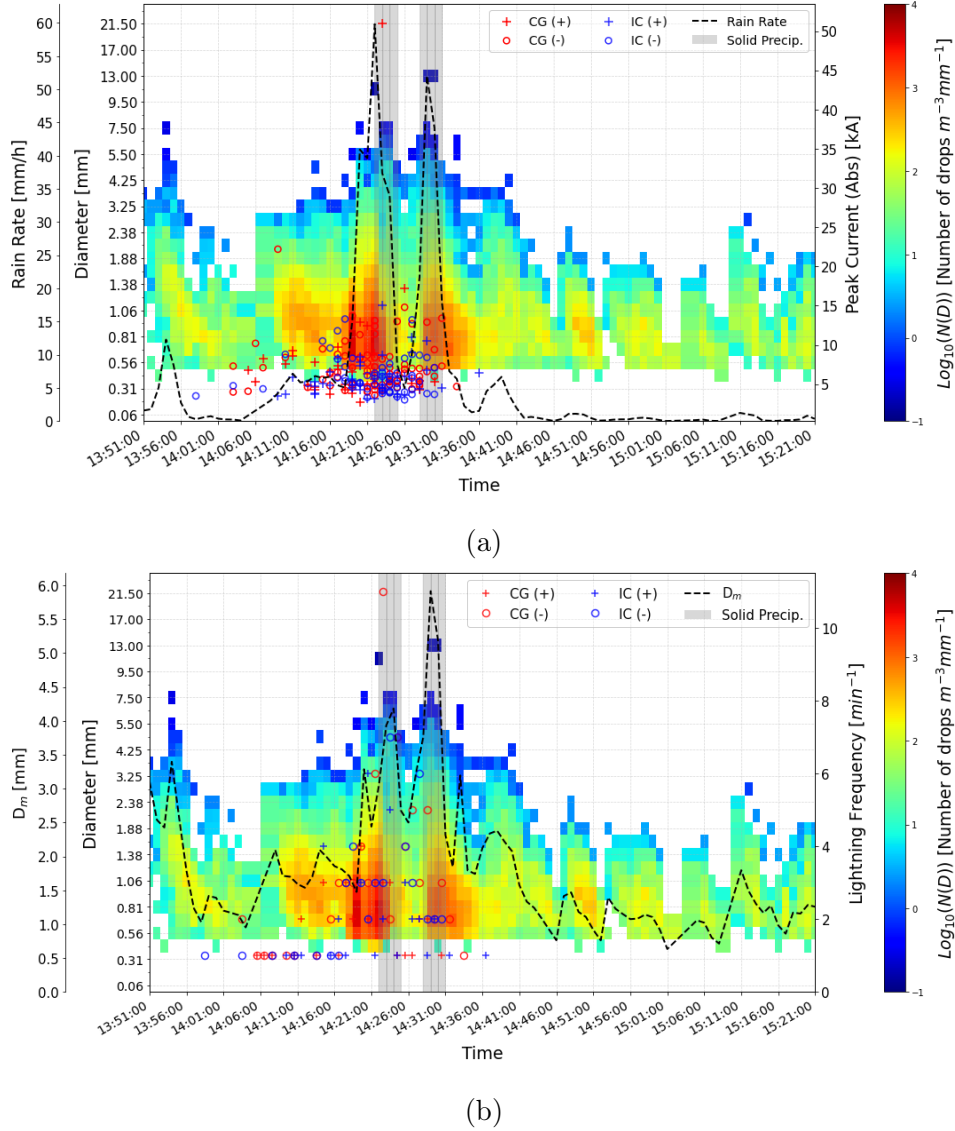


Figure 4.12: Temporal evolution of the logarithmic drop size distribution for Ev. 17. Panel (a) shows the RR profile overlaid with the absolute peak current of lightning strokes. Panel (b) displays the D_m overlaid with the lightning stroke frequency (strokes min⁻¹). In both panels, lightning activity within a 5 km radius is categorized by discharge type and polarity.

The convective event, as recorded by the disdrometer, lasts approximately 90 minutes,

and lightning activity is not detected during its initial and final stages, as shown in Figure 4.12. As usual, the onset of the storm is characterized by large hydrometeors, reaching a maximum diameter of 4.0 mm, while smaller drops are completely absent. In the subsequent minutes, larger raindrops with diameters up to 7.5 mm are detected, alongside an increase in small-drop concentrations. This initial phase, lasting about 15 minutes, is succeeded by the actual mature phase of the storm characterised by intense lightning activity. Here, hydrometeors progressively increase their dimensions, hailstones are detected peaking at a diameter of 13.0 mm, while small-drop concentrations gradually rise to their highest recorded values during the event, in the diameter range 0.5 – 1.88 mm resulting in two RR peaks (Figure 4.12a). Between these two peaks, which are separated by a 7-minute interval, a temporary decline in precipitation intensity is observed, driven by an overall reduction in raindrop concentrations; this leads to a sharp drop in RR value. Interestingly, the detection of hail ceases for only 3 minutes before resuming in correspondence with a second and higher D_m peak, capped at around 6.0 mm. The explanation for this brief interruption in solid precipitation and reduction in overall precipitation intensity lies in the complex structural organization of the storm cell during its transit. This dynamic will be investigated through the spatial analysis of the radar data in the upcoming sections.

During the mature phase, a high number of lightning strokes is detected, with frequency peaking at approximately 11 strokes min^{-1} (Figure 4.12b). These strokes are predominantly clustered where solid precipitation is present. Within this cluster, the most intense stroke (CG), reaching a peak current of 50 kA, occurs exactly where both hail and the largest raindrops are recorded. Following the hailfall, the electrical activity undergoes a rapid decline, consistent with the expectation that the fallout of hail effectively discharges the cloud by acting as a carrier for downward charge transport. The stroke rate drops to near-unit values before vanishing completely as the storm enters its final stage marked by a gradual decrease in overall concentrations and raindrop diameters, now peaking at around 3.0 mm. In this weakening phase, which lasts approximately 40 minutes, the precipitation assumes a stratiform nature, characterized by the absence of significant peaks in the RR and D_m values, which both exhibit relatively flat profiles.

4.2.2 Phase Segmentation and Electrical Trends

To evaluate the electrical response of the storm cloud to solid precipitation, each convective episode is divided into three distinct phases. Utilizing ground-based observations, the three specific temporal windows are defined as follows: a 5-minute phase prior to the onset of the hailfall (*Pre-Hail*), a phase coinciding with the solid precipitation (*During-Hail*), and a trailing phase spanning the 5 minutes after the cessation of the hailfall (*Post-Hail*).

For each of these phases, two metrics are extracted to quantify the evolution of electrical activity: the average peak current (AC) and the percentage distribution of stroke counts. Both metrics are assessed for the overall lightning activity as well as for its breakdown into intra-cloud and cloud-to-ground discharges.

The average peak current intensity (in kA), calculated as the absolute value for all discharges within each window, is reported in Table 4.7 and illustrated in Figure 4.14. For

the stroke fraction analysis (Table 4.6 and Figure 4.13), a standardization procedure is required. Since the *During-Hail* phase exhibits a variable duration from event to event, the counts recorded during the core transit are normalized to a virtual 5-minute interval. The percentage fractions reported in the plots are then calculated by defining the ratio of the strokes in each phase to the overall event count of its respective category, based on the sum of the three windows of equal virtual duration. This temporal homogenization of the three windows allows for a direct comparison across the phases.

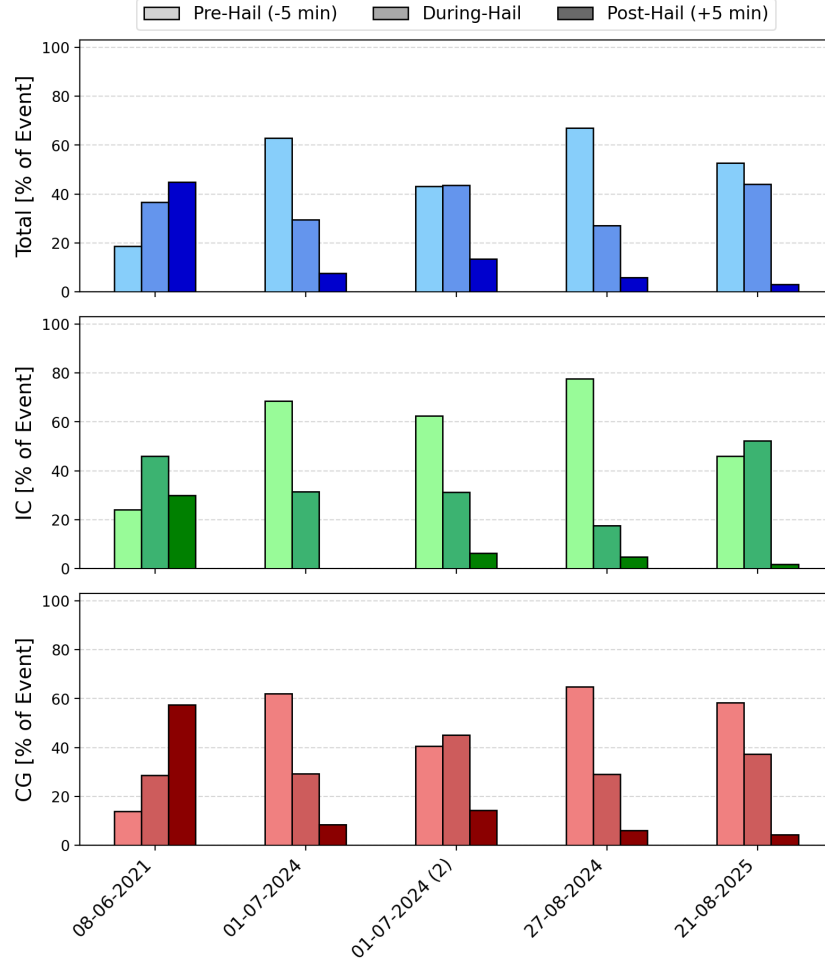


Figure 4.13: The bar plots show the lightning stroke fractions (%) across three temporal phases of hail fallout for the five analyzed events. The 'During-Hail' counts are normalized to a 5-minute interval to ensure consistency with the Pre-Hail (-5 min) and Post-Hail (+5 min) windows. Panels show the stroke distribution for Total (top), intra-cloud (middle), and cloud-to-ground (bottom) categories.

As previously observed in the microphysical analysis derived from the Parsivel² disdrometer data, a consistent trend is immediately visible in Figure 4.13 for four out of the five selected cases (01-07-2024, 01-07-2024 (2), 27-08-2024, and 21-08-2025). For these events, the highest stroke fractions occur in the 5 minutes preceding the onset of solid precipitation, with Total stroke percentages ranging from 43% to 67% (Figure 4.13, top panel). This initial accumulation is particularly evident in the intra-cloud discharges, which peak

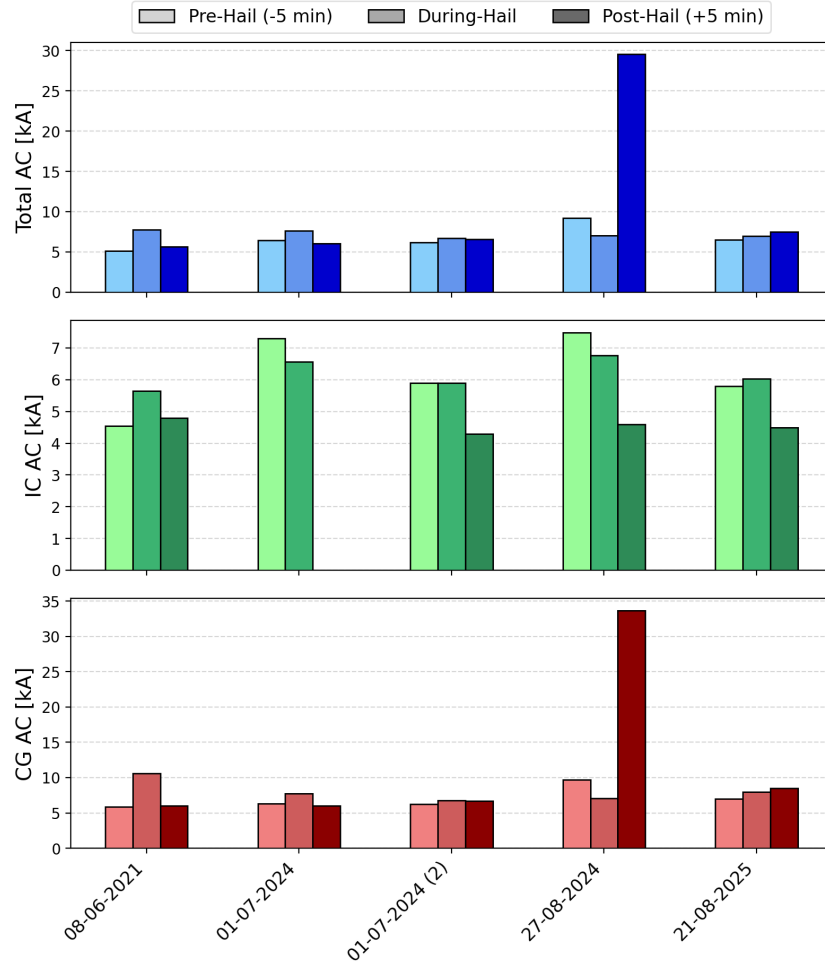


Figure 4.14: The bar plots show the average peak current intensity (kA) of lightning strokes across three temporal phases of hail fallout for the five analyzed events. The Pre-Hail and Post-Hail windows are defined as 5-minute intervals relative to the hail core passage. Panels show the Average Current (AC) for Total (top), intra-cloud (middle), and cloud-to-ground (bottom) categories.

at 68.5% for the Ev. 09 and reach an impressive 77.5% during the Ev. 12 (Figure 4.13, central panel). As the storm transitions into the *During-Hail* phase, a marked decrease in the normalized lightning fraction is observed. This reduction becomes drastic in the *Post-Hail* phase, where the electrical activity collapses. For instance, the Total stroke fractions drop to marginal values such as 7.6% for Ev. 09, 5.8% for Ev. 12, and 3.1% for Ev. 17. The depletion is even more extreme for the IC category, which falls to 1.7% for Ev. 17, and completely ceases (0.0%) in the first 01-07-2024 case. The intense electrical activity observed in the *Pre-Hail* window indicates the presence of a mature and charged convective core, where electrical charges are effectively separated and accumulated in stratified regions aloft within the cloud. The subsequent and progressive decline of the lightning fractions during the hailfall and finally in the *Post-Hail* phase indicates that the precipitating solid hydrometeors are carrying these charges downward, disrupting the stratified charge centers and leading to the electrical discharge of the cloud. These findings provide evidence that hail effectively acts as a carrier for downward charge trans-

port.

As previously observed, the 08-06-2021 event exhibits an opposite trend compared to the other analyzed cases. In this instance, the Total electrical activity increases from 18.6% in the *Pre-Hail* phase to a peak of 44.8% in the *Post-Hail* phase, driven largely by a surge in CG strokes reaching 57.4% (Figure 4.13, bottom panel). This divergence highlights a distinct structural evolution of this specific event and an investigation of the storm's organization and its radar reflectivity patterns will be addressed in the following sections to better characterize this behavior.

Table 4.6: Lightning stroke distribution fractions (%) across hail phases.

ID / Date	Phase	Total (%)	IC (%)	CG (%)
Ev. 08 / 08-06-2021	Pre-Hail	18.6	24.2	13.9
	During-Hail	36.6	45.9	28.7
	Post-Hail	44.8	30.0	57.4
Ev. 09 / 01-07-2024	Pre-Hail	62.8	68.5	62.1
	During-Hail	29.6	31.5	29.3
	Post-Hail	7.6	0.0	8.6
Ev. 10 / 01-07-2024 (2)	Pre-Hail	43.1	62.5	40.5
	During-Hail	43.5	31.2	45.2
	Post-Hail	13.4	6.2	14.3
Ev. 12 / 27-08-2024	Pre-Hail	67.1	77.5	64.9
	During-Hail	27.1	17.6	29.1
	Post-Hail	5.8	4.8	6.0
Ev. 17 / 21-08-2025	Pre-Hail	52.8	46.1	58.4
	During-Hail	44.1	52.2	37.4
	Post-Hail	3.1	1.7	4.3

Note: All phases represent a 5-minute time window. The 'During-Hail' phase has been normalized to a 5-minute equivalent duration to allow direct comparison with the Pre-Hail and Post-Hail phases.

While stroke counts indicate a clear decreasing trend, the dynamics of the discharge intensity provide further insight as the average peak current intensity exhibits a more resilient behavior across the three phases (Figure 4.14). In the four convective cases, the current intensity remains relatively stable or undergoes only moderate fluctuations during the *During-Hail* phase. This observation is particularly significant: while the number of discharges drops drastically, the residual electrical activity maintains an intensity comparable to the *Pre-Hail* phase. An exception is represented by the 27-08-2024 event, which exhibits a substantial increase in average current during the *Post-Hail* phase, where the intensity rises from approximately 9 and 7 kA in the *Pre-Hail* and *During-Hail* windows, respectively, to 29.53 kA for the Total count. This increase is heavily

driven by the CG discharges, which record an average current of 33.68 kA during this phase. As previously observed in the heatmaps, this surge is attributable to an isolated, high-intensity positive cloud-to-ground discharge occurring in the final minutes of the convective phase, peaking at about 140 kA (Figure 4.11a and Table 4.5). The occurrence of this single extreme event skews the CG average, elevating the Total average current for the entire window.

Table 4.7: Average peak current intensity (kA) per lightning type across hail phases.

ID / Date	Phase	Total (kA)	IC (kA)	CG (kA)
Ev. 08 / 08-06-2021	Pre-Hail	5.10	4.55	5.90
	During-Hail	7.74	5.64	10.59
	Post-Hail	5.66	4.80	6.04
Ev. 09 / 01-07-2024	Pre-Hail	6.46	7.30	6.35
	During-Hail	7.63	6.56	7.78
	Post-Hail	6.05	0.00	6.05
Ev. 10 / 01-07-2024 (2)	Pre-Hail	6.20	5.90	6.26
	During-Hail	6.71	5.90	6.79
	Post-Hail	6.56	4.30	6.69
Ev. 12 / 27-08-2024	Pre-Hail	9.24	7.49	9.67
	During-Hail	7.03	6.76	7.07
	Post-Hail	29.53	4.60	33.68
Ev. 17 / 21-08-2025	Pre-Hail	6.54	5.80	7.03
	During-Hail	6.94	6.03	7.99
	Post-Hail	7.48	4.50	8.47

Note: Phases are defined as 5-minute intervals relative to the hail core passage.

4.3 3-D Radar Diagnostics and Phenomenological Coupling

The characterization of precipitating hydrometeors at the surface provides crucial insights into the localized fallout phase of a storm. However, since a ground-based disdrometer exclusively records the particles that survive the fall below the cloud base, it lacks the capability to observe the vertical structure of the convective cell and the internal updraft dynamics. A purely ground-level perspective is insufficient to fully characterize the convective phenomenon. To bridge this observational gap and establish a coupling between the conditions observed at the surface and those occurring aloft, this section employs the 3-D volumetric data collected by the ARPAE weather radar network, utilizing the San Pietro Capofiume and Gattatico installations. This approach allows for the reconstruction and analysis of the vertical structure of the storm.

Furthermore, since the laser disdrometer performs a strictly point-like measurement at the ground, sampling hydrometeors crossing a sensitive area of just 54 cm^2 , it cannot, on its own, describe the full spatial evolution of a storm. To overcome this limitation, radar maps are employed to provide a macroscopic visualization of the convective system, enabling the tracking of its horizontal extent and overall trajectory.

Beyond the constraints of spatial representativeness, the integration of such heterogeneous sensors introduces an additional challenge due to their deeply asynchronous sampling frequencies. The precipitation spectrum from the disdrometer is continuously recorded with an integration time of 1 minute, and the LINET network records lightning strokes in real-time, down to the microsecond. On the other hand, the ARPAE weather radars operate on much longer time scales, completing scanning volumes every 5 minutes during severe weather conditions, as detailed in Section 2.4. This sampling asynchrony implies that between consecutive radar acquisitions, an observational gap of 5 minutes occurs. Within this temporal window, both the electrical dynamics and the ground microphysics can exhibit abrupt variations and transient activity peaks.

Accounting for these observational and temporal constraints, the subsequent analysis is not systematically carried out for all selected events. Instead, it is focused on specific case studies where ground-level disdrometer observations alone are insufficient to fully unravel the underlying convective dynamics and the integration of 3-D radar fields becomes essential to resolve apparent microphysical mismatches and reconstruct the volumetric evolution of the storm.

4.3.1 Spatial Evolution and Regional Storm Dynamics

To overcome the spatial limitations inherent in the point-based measurements of the disdrometer, the spatial evolution of the storm cells is investigated through composite radar reflectivity maps. The data processing and the methodology applied for the generation of these composite fields have been extensively detailed in Section 3.3. The spatial domain of the maps encompasses a regional area bounded between 9.8°E and 12.8°E in longitude, and 43.6°N and 45.4°N in latitude. Within this domain, the geographical positions of the observational instruments are marked: the Parsivel² disdrometer (44.4993°N , 11.3538°E), alongside the two regional radar stations of Gattatico (GAT) (44.7914°N , 10.4992°E) and San Pietro Capofiume (SPC) (44.6547°N , 11.6236°E) utilized to compute the composite fields.

By tracking the structural organization and the trajectory of the convective cores across this domain, it is possible to contextualize the microphysical anomalies and the temporal gaps previously observed in the disdrometer time series for the selected events. This spatial perspective allows for a broader understanding of the storm dynamics that cannot be fully appreciated through isolated ground-based observations.

Ev. 08 - June 8th, 2021

The event of June 8th, 2021, is characterized by a linear mesoscale organization. Specifically, the composite radar reflectivity maps display the passage of a squall line, which consists of convective clouds organized in a linear fashion, featuring a leading convective edge followed by a wide stratiform area [9]. Consistent with this macro-structure, the radar fields reveal a leading convective line containing a well-defined high-reflectivity

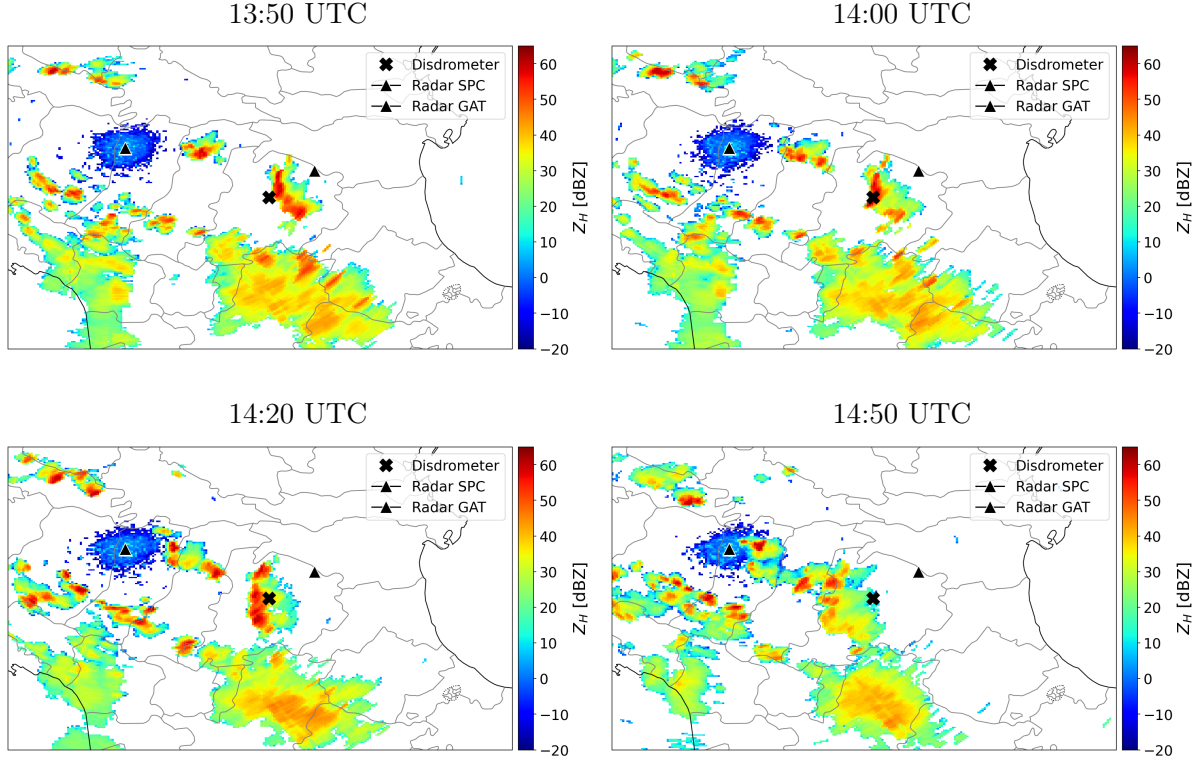


Figure 4.15: Temporal evolution of composite radar reflectivity (Z_H) for Ev. 08 (June 08, 2021). The panels show the storm’s propagation relative to the disdrometer (black cross) and the reference locations of the SPC and GAT radar stations (black triangles).

core, with Z values exceeding 55 dBZ and a very steep spatial gradient, as shown in Figure 4.15.

Specifically, the first map illustrates the conditions at 13:50 UTC, showing the convective system propagating from Northeast to Southwest and approaching the ground station. The map corresponding to 14:00 UTC captures the moment of maximum impact, when the precipitation core is positioned over the sampling site. High reflectivity values are detected in this phase, which is corroborated by the disdrometer simultaneously recording solid precipitation (Figure 4.8). Starting from 14:20 UTC, the disdrometer registers a substantial decrease in precipitation intensity. This drop coincides with the passage of the lower-reflectivity trailing region (around 30 dBZ) of the convective system over the sensor, characterizing the stratiform precipitation phase. Finally, in the radar map at 14:50 UTC, the storm has moved away from the observation area, leaving the disdrometer site completely clear of radar echoes and marking the definitive conclusion of the precipitation event.

By relating these radar observations to the electrical activity recorded by the LINET network, the lightning rates observed in Figure 4.8 after the cessation of solid precipitation find an explanation. While the narrow convective core responsible for the localized hail fallout rapidly passes over the point-like disdrometer, the broader convective structure of the squall line continues to propagate across the domain. Consequently, the LINET network continues to detect intense electrical activity generated by adjacent convective clouds within the squall line.

Ev. 12 - August 27th, 2024.

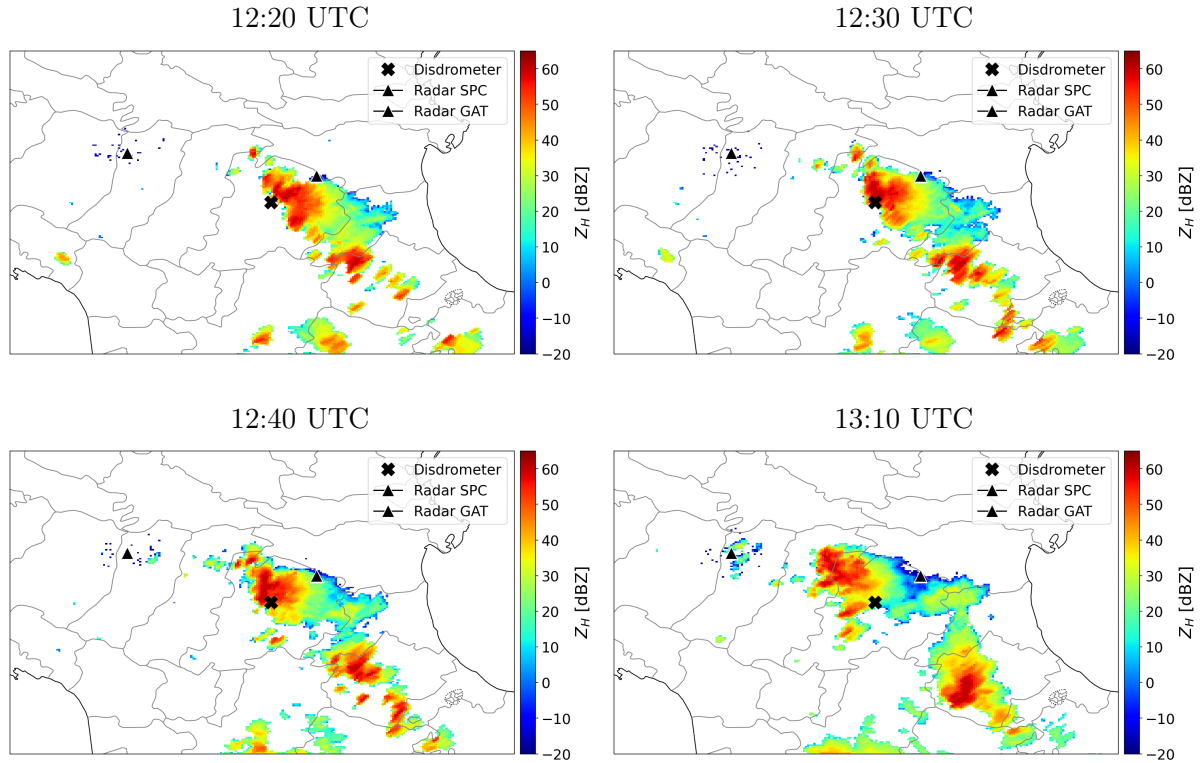


Figure 4.16: Temporal evolution of composite radar reflectivity (Z_H) for Ev. 12 (August 27, 2024). The panels show the storm’s propagation relative to the disdrometer (black cross) and the reference locations of the SPC and GAT radar stations (black triangles).

The precipitation event of August 27th, 2024, consists of a multicellular cluster characterized by an East-Southeast to West-Northwest propagation (ESE-WNW). The radar map sequence from 12:20 UTC to 13:10 UTC documents the slow approach of the convective cell, its prolonged stationary phase over the disdrometer site, and its final departure.

In the radar map relative to 12:20 UTC, the measurement site is still clear of precipitation, while intense convective cells are active to the East and Southeast and the system has experienced a significant westward expansion and propagation toward the station. At 12:30 UTC, the severe core of the system, featuring maximum reflectivity values exceeding 60 dBZ, is positioned directly above the disdrometer. While intense electrical activity is detected by the LINET network at this time, confirming the storm’s approach, the Parsivel² disdrometer does not record any hydrometeors until 12:37 UTC (Figure 4.11). As previously detailed in Section 4.2.1, this absence of data is due to a technical initialization delay in the instrument’s data logging system, rather than a lack of precipitation at the surface.

The convective core maintains its peak intensity over the sampling site for a prolonged period: at 12:40 UTC high reflectivity values are still measured, corresponding to the detection of hail at the surface (Figure 4.11). The system remains virtually stationary above the disdrometer for approximately 20 minutes, resulting in precipitation of lower intensity, until 13:10 UTC. The radar map at this time shows the high-reflectivity core shifting towards the Northwest, leaving the measurement area free of precipitation.

Ev. 17 - August 21st, 2025.

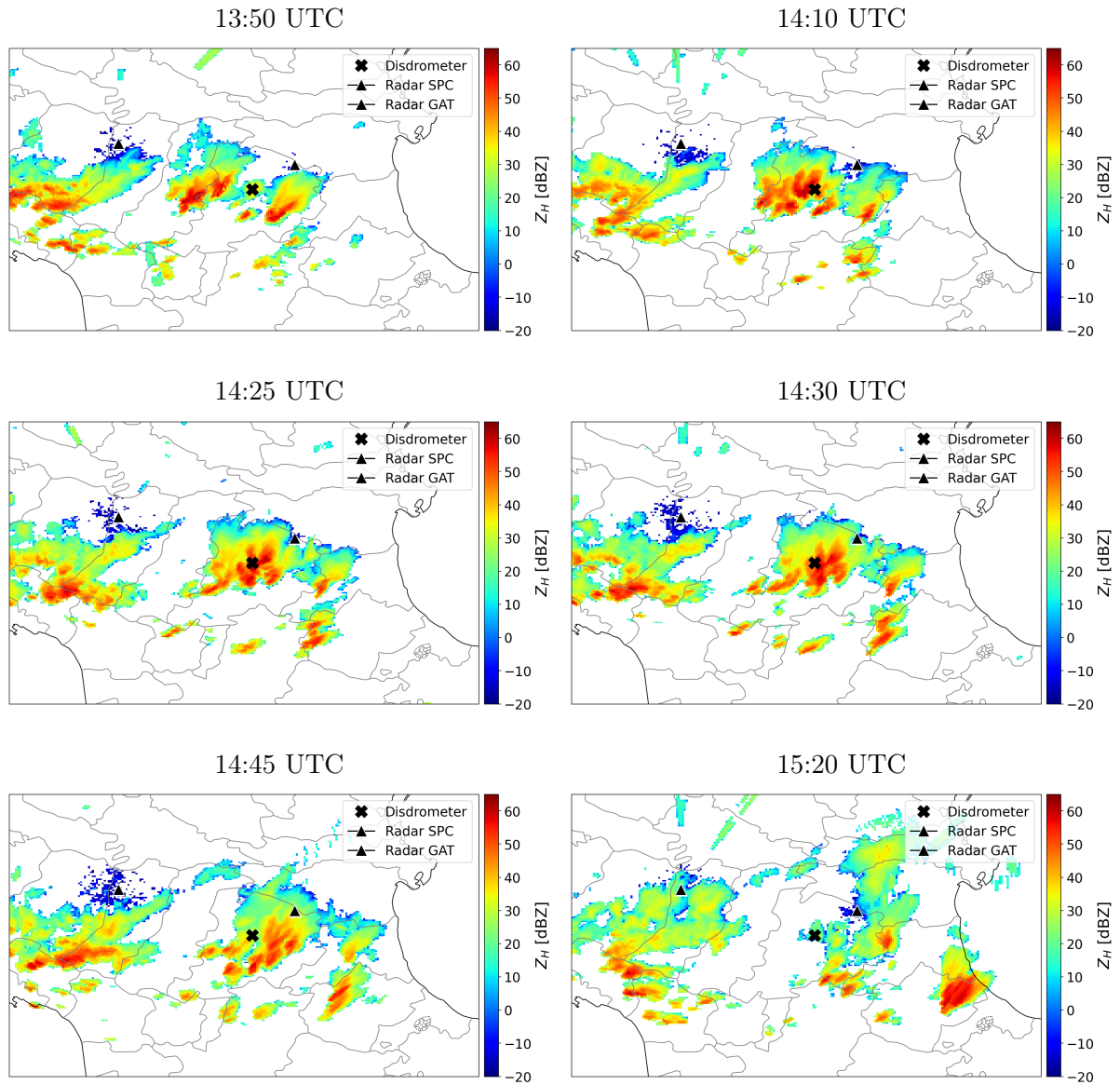


Figure 4.17: Temporal evolution of composite radar reflectivity (Z_H) for Ev. 17 (August 21, 2025). The panels show the storm's propagation relative to the disdrometer (black cross) and the reference locations of the SPC and GAT radar stations (black triangles).

The precipitation event recorded on August 21st, 2025 is associated with a mesoscale multicellular system originating west-southwest of the disdrometer's location. These active convective cells merge and organize along a front, following a clear trajectory that propagates toward the east-northeast (WSW-ENE), bringing sustained and intense precipitation over the target domain.

This event exhibits a more complex dynamic compared to the previously analyzed case studies. At the onset of the precipitation record, the disdrometer detects low-intensity precipitation, which is consistent with the moderate reflectivity core positioned over the measurement site in the 13:50 UTC map, showing Z_H values peaking at 35 dBZ. In the 14:10 UTC map, a northern convective cell merges with the pre-existing system,

forming a unified cluster that continues to propagate in the east-northeast direction. Concurrently, the surface sensor detects a marked intensification in the precipitation rate, accompanied by the initiation of electrical activity.

At 14:25 UTC, the convective core, featuring reflectivity values exceeding 50 dBZ, envelopes the measurement site and the disdrometer, indeed, detects solid precipitation at the surface. However, as previously reported in Section 4.2.1, the concentration spectra reveal a brief interval of a few minutes (between 14:25 UTC and 14:28 UTC) separating two distinct hail phases. During this gap, the precipitation intensity drops, characterized by a general decrease in both the size and concentration of the recorded hydrometeors. This surface observation is corroborated by the radar kinematics between 14:25 UTC and 14:30 UTC, suggesting that a sector of the storm with lower reflectivity, presumably the edge of the convective core, transits over the site and causes the temporary decrease in surface rain rate. At 14:30 UTC, the highest reflectivity core sweeps back over the disdrometer, which coincides with the return of solid precipitation. As shown in the 14:45 UTC map, the most active and severe portion of the thunderstorm leaves the measurement site, propagating further to the east. This shift leads to a decrease in precipitation intensity over the disdrometer, which is now covered by the trailing region of the storm, characterized by residual precipitation and weaker radar echoes ranging between 20 and 30 dBZ. Finally, the 15:20 UTC map illustrates the concluding phase of the event: the main system has definitively cleared the disdrometer and considerably weakened further east.

4.3.2 Vertical Cross-Sectional Analysis and Electrical Response

As previously discussed, the disdrometer at the surface provides no insight into the storm’s vertical development or the processes evolving aloft. To address this limitation, volumetric radar data are employed to extract pseudo-Range Height Indicator (RHI) cross-sections of horizontal radar reflectivity (Z_H), following the methodology detailed in Section 3.3. Data are primarily sourced from the San Pietro Capofiume radar station, which is selected due to its proximity to the disdrometer site, thereby minimizing signal attenuation. For the specific case of the June 8th, 2021 event, data from the Gattatico radar are utilized instead, owing to the temporary unavailability of the SPC station during that period.

Thus, the following section focuses on the pseudo-RHI analysis, enabling a detailed inspection of the convective core structure and its temporal evolution in relation to lightning activity. Specifically, the cross-section spans a 10 km horizontal window, centered on the Parsivel² disdrometer location, which acts as the fixed point for the azimuthal cut. The specific azimuthal angle of the cut is not fixed, but is selected to intersect the storm either parallel or perpendicular to its direction of motion, which is derived from the spatial tracking of the radar maps. The horizontal axis of these plots displays the geographical coordinates, latitude and longitude, and the distance from the radar in kilometers, while the vertical axis represents the height in kilometers above ground level (AGL). Lightning data from the LINET network, occurring within a 5 km horizontal radius of the disdrometer, are projected onto the radar pseudo-RHI plots, with all discharges distinguished by polarity. IC flashes are plotted as scatter points corresponding to their emission heights, whereas CG strokes are visualized in a dedicated lower panel

as histograms, where the number of events is aggregated into 0.5 km spatial intervals along the azimuthal cross-section.

Ev. 08 - June 8th, 2021.

Figure 4.18 shows the evolution of the precipitating core for the June 8th, 2021 event, through a sequence of pseudo-RHI cross-sections extracted at a fixed azimuthal angle of 225°. This orientation is parallel to the storm's propagation trajectory (from NE to SW), providing a longitudinal slice that captures the advection and the internal dynamics of the squall line as it transits over the observation site.

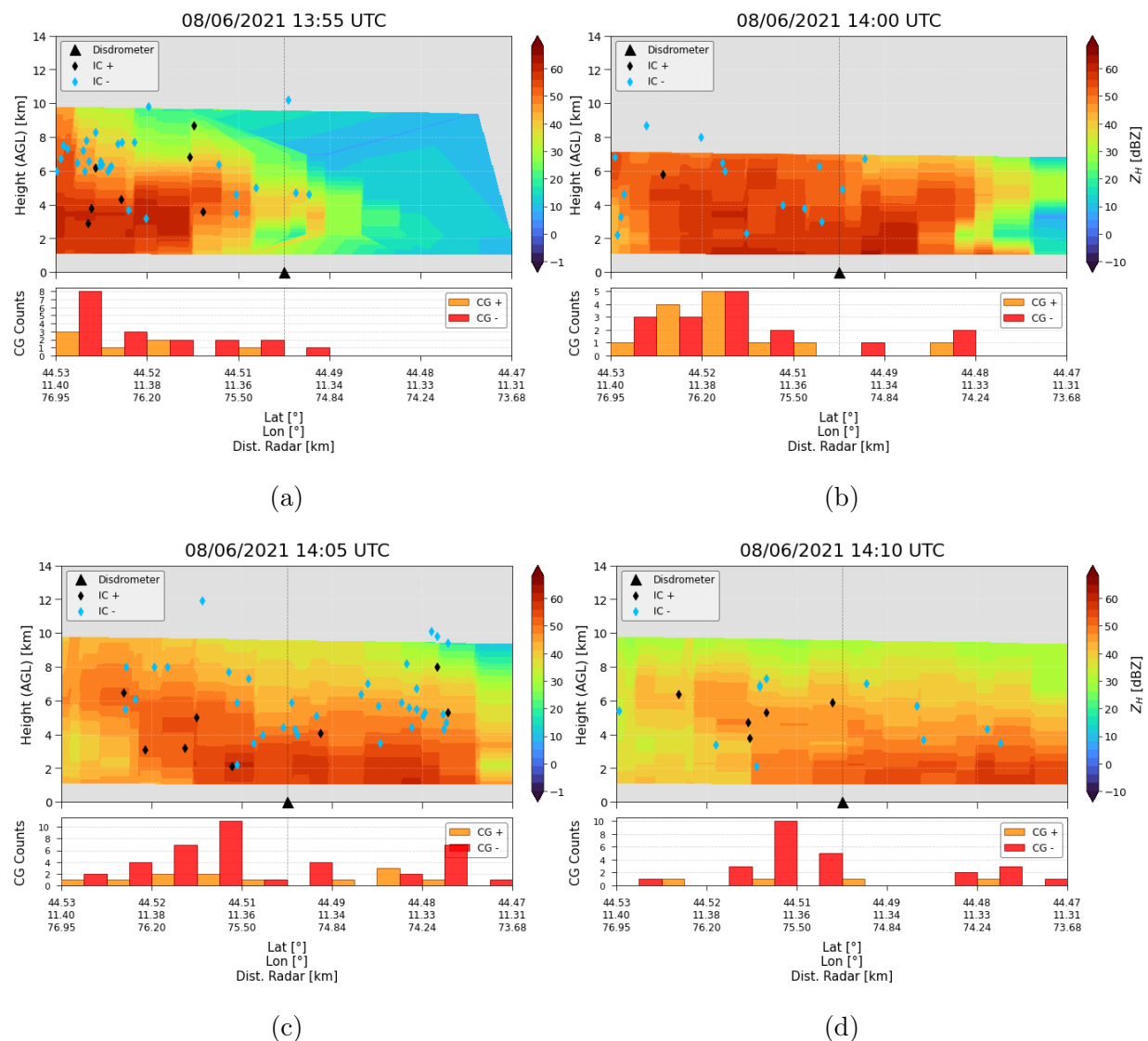


Figure 4.18: Temporal evolution of the convective system for Ev. 08 (June 08, 2021). The upper panels show the Z_H cross-sections, while the lower panels display the spatial occurrence of CG strokes, aggregated into 0.5 km intervals along the azimuthal cross-section.

At 13:55 UTC, the system is approaching from the northeast (Figure 4.18a). The cross-

section reveals an intense convective core with maximum Z_H values exceeding 60 dBZ, extending vertically up to 10.0 km. The electrical activity, characterized by a dense mixture of IC and CG strokes, is heavily concentrated within this advancing leading edge. By 14:00 UTC (Figure 4.18b), the high-reflectivity core reaches the disdrometer. The vertical profile reveals the progressive descent of the highest reflectivity values toward the lower levels. Coincident with this descent, the disdrometer actively detects the onset of solid precipitation at the surface. Concurrently, there is an increase in CG stroke rates, particularly of negative polarity, spatially collocated with this descending mass, while IC flashes also exhibit a clear predominance of negative discharges.

The most intense phase of the event regarding the electrical activity occurs at 14:05 UTC, as shown in Figure 4.18c. The cross-section highlights a wide area extending from the upper levels down to the surface, characterized by intense precipitation. This convective structure coincides with a peak in negative CG activity, with a maximum occurrence of 10 strokes per 0.5 km spatial interval. However, as already noted in Section 4.2.1, this electrical response is not strictly confined to the local convective core; as an inherent feature of the squall line's broader structure, the surrounding environment and adjacent cells contribute significantly to the observed lightning activity, reflecting the collective electrical behavior of the propagating system.

Finally, the radar scan at 14:10 depicts the advection phase of the system as it translates further southwest, moving away from the sensor (Figure 4.18d). As the main convective core shifts, the surface reflectivity above the disdrometer drops, corresponding to a decrease in detected precipitation at the ground site. The spatial distribution of the lightning strikes tracks the horizontal advection of the high-reflectivity core; as the local electrical activity over the sensor collapses, this indicates a strong correlation between the passage of the intense convective core and the occurrence of electrical discharges.

Ev. 12 - August 27th, 2024.

The convective event of August 27th, 2024, is documented in Figures 4.19 and 4.20, which illustrate its temporal evolution between 12:25 and 12:45 UTC. The analysis utilizes a dual perspective: longitudinal pseudo-RHI cross-sections extracted along the storm's direction of propagation (from ESE to WNW) at an azimuthal angle of 315°, and transverse cross-sections extracted perpendicularly at 45°.

During the initial approaching phase at 12:25 UTC (Figure 4.19a), the longitudinal cross-section reveals an advancing convective system characterized by a broad and elevated high-reflectivity region. At this stage, maximum Z_H values are predominantly in the 45 – 55 dBZ range, extending vertically well beyond 12 km. The electrical activity is already significant, with numerous IC flashes scattered across the upper levels of the storm, the majority occurring between 8 and 14 km, and a considerable number of CG strokes recorded at the surface.

By 12:30 UTC (Figure 4.19b), the system reaches its maximum structural maturity, featuring a distinct high-reflectivity nucleus suspended between 4 and 11 km almost directly above the disdrometer, with Z_H reaching 60 dBZ. The same high-intensity echo is observed in the transverse cut at the same timestamp (Figure 4.20a). This cross-section reveals a pronounced inclination of the convective structure. Combined with the Weak Echo Region (WER) observed at lower altitudes near the surface and the extremely high echo tops, exceeding 12 km, these signatures are consistent with the morphology of

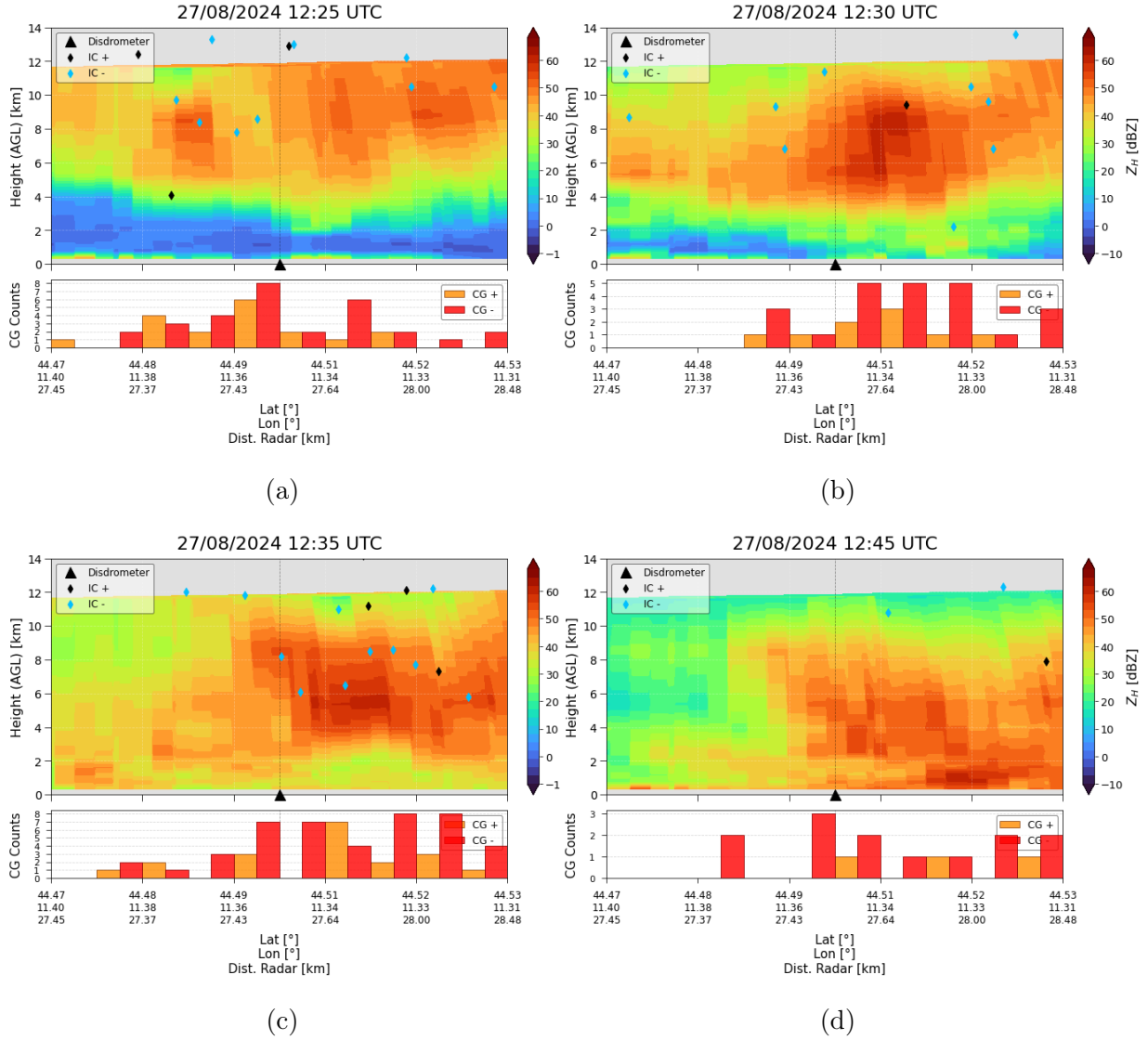


Figure 4.19: Temporal evolution of the convective system for Ev. 12 (August 27, 2024). The upper panels show the Z_H cross-sections, while the lower panels display the spatial occurrence of CG strokes, aggregated into 0.5 km intervals along the azimuthal cross-section.

supercellular storms, driven by a strong updraft that maintains the high-reflectivity core in suspension.

The subsequent 12:35 UTC timestamp (Figure 4.19c) captures a drastic structural evolution. Once the accumulated mass of solid precipitation outweighs the vertical support of the updraft, the overhanging structure undergoes a rapid collapse. The high- Z_H core descends toward the surface, entirely filling the previously observed WER. Despite the disdrometer's activation delay, as stated in Section 4.2.1, the instrument records strong precipitation at the ground starting at 12:37 UTC, with solid hydrometeors detected from 12:40 UTC onwards. During this phase, the storm exhibits its peak in electrical activity: a cluster of IC flashes is localized precisely within the suspended nucleus, while high frequencies of CG strokes are detected directly beneath it, within the 0.5 km spatial

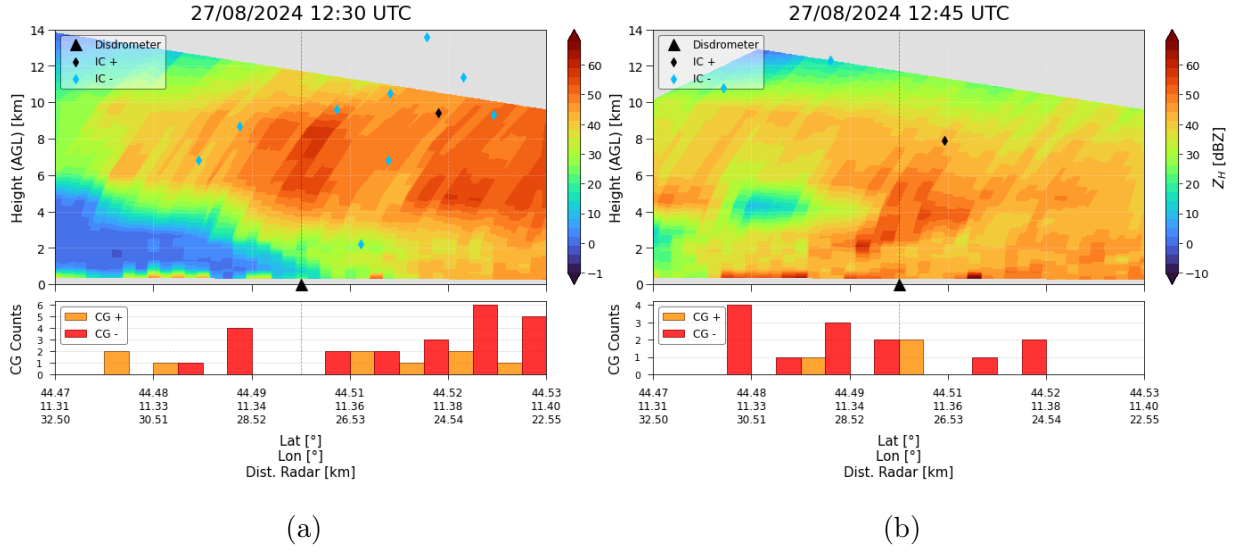


Figure 4.20: Temporal evolution of the convective system for Ev. 12 (August 27, 2024). The upper panels show the transversal Z_H cross-sections, while the lower panels display the spatial occurrence of CG strokes, aggregated into 0.5 km intervals along the azimuthal cross-section.

interval.

By the final 12:45 UTC scan, shown in Figure 4.19d and 4.20b, while the ground-based sensor continues to record intense precipitation, the higher reflectivity values are now strictly confined to the ground level, and the overall lightning activity undergoes a drastic reduction, validating the role of descending solid precipitation in discharging the convective core.

Ev. 17 - August 21st, 2025.

The convective event of August 21st, 2025 is reported in Figure 4.21, which illustrates the temporal evolution of the complex system through a four-panel sequence of pseudo-RHI cross-sections (14:10, 14:15, 14:20, and 14:25 UTC). Consistent with the analytical approach used for the previous event, these cross-sections are extracted along the storm's direction of motion (from WSW to ENE) at an azimuthal angle of 67°.

During the approaching phase at 14:10 and 14:15 UTC (Figure 4.21a, 4.21 b), the advancing convective core exhibits a high-reflectivity column reaching extremely high values, up to 65 dBZ in the lower levels, and extending vertically up to 9 km AGL. The electrical activity progressively intensifies, characterized by an increase in both the number of detected IC flashes and the occurrence of CG strokes, while maintaining a strict spatial alignment with the advancing high- Z_H core.

The storm reaches its maximum electrical intensity at 14:20 UTC (Figure 4.21c), with an increase in both IC flashes and CG strokes, with high frequencies concentrated around the ground-based sensor. Surface disdrometer observations indicate that the fallout of solid hydrometeors begins precisely at 14:22 UTC (Figure 4.12), concurrently with this peak in lightning activity. In the transition between 14:15 and 14:20 UTC, the convective system reveals a complex internal structure, likely resulting from a multi-cellular

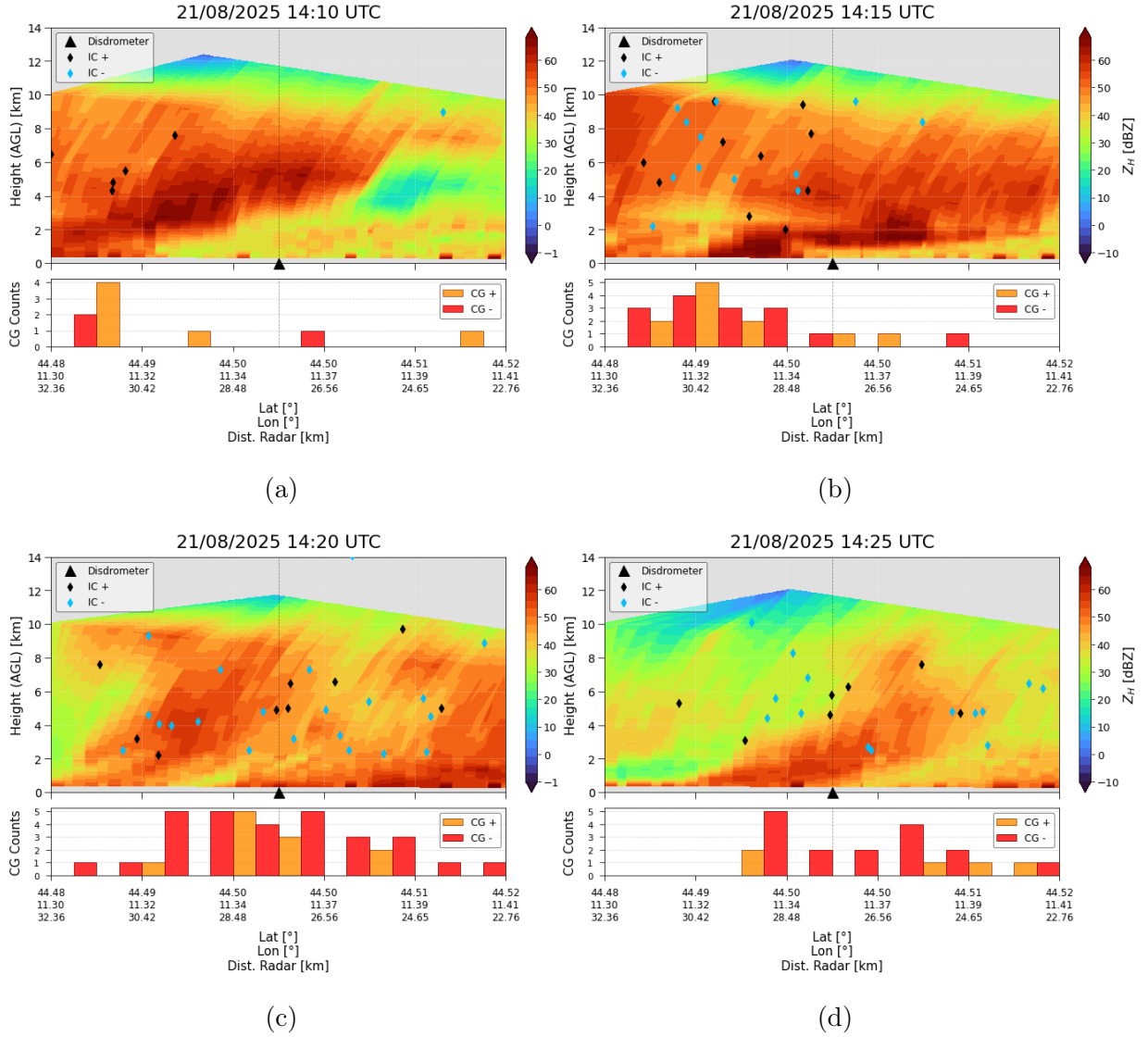


Figure 4.21: Temporal evolution of the convective system for Ev. 17 (August 21, 2025). The upper panels show the Z_H cross-sections, while the lower panels display the spatial occurrence of CG strokes, aggregated into 0.5 km intervals along the azimuthal cross-section.

configuration, evidenced by adjacent high-reflectivity columns passing sequentially over the disdrometer. This structural feature is clearly visible in transverse pseudo-RHI cross-sections, extracted perpendicularly to the direction of propagation of the storm at a fixed azimuthal angle of 337° . At the 14:20 UTC timestamp (Figure 4.22b), two distinct nuclei are discernible: a high-intensity one located to the left and a weaker one to the right of the disdrometer position. Together with the 14:15 UTC scan (Figure 4.22a), these perpendicular cuts reveal a striking feature: localized clusters of IC flashes are strictly confined within the convective nucleus of the thunderstorm characterized by high Z_H values, with CG strokes also concentrated beneath it. Given that surface disdrometer observations confirm the presence of solid precipitation, the electrical discharges appear "caged" within the vertical channel where the solid mass concentration is at its highest.

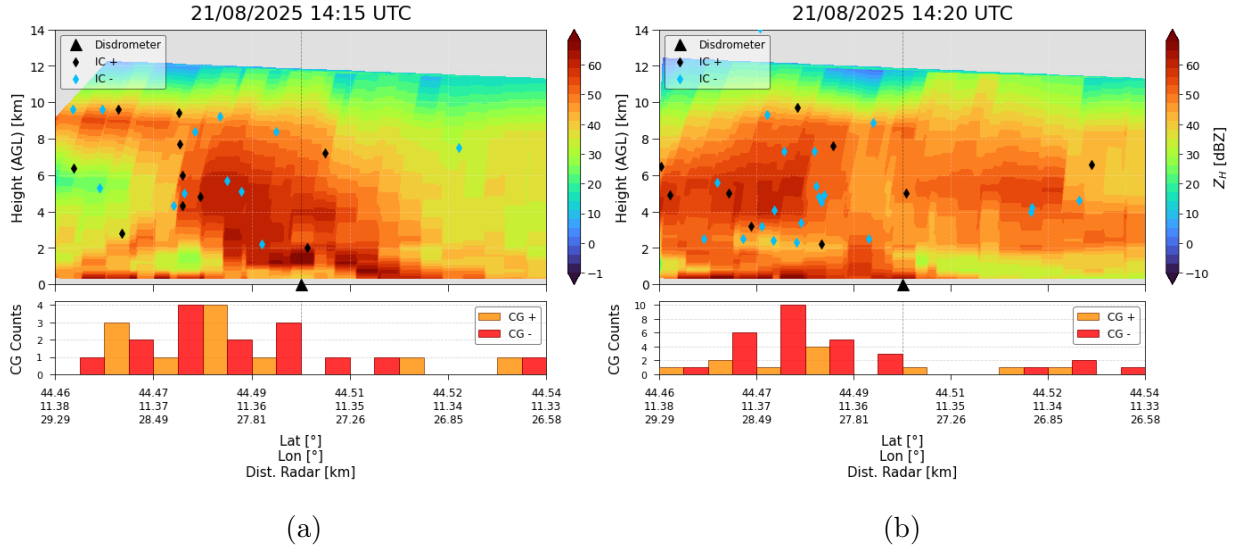


Figure 4.22: Temporal evolution of the convective system for Ev. 17 (August 21, 2025). The upper panels show the transversal Z_H cross-sections, while the lower panels display the spatial occurrence of CG strokes, aggregated into 0.5 km intervals along the azimuthal cross-section.

By the 14:25 UTC scan (Figure 4.21d), the high reflectivity values are present almost exclusively at the ground, where the disdrometer still records the continuous fallout of solid precipitation, while the electrical activity undergoes a reduction. The decline in lightning frequency, occurring concurrently with the solid phase reaching the surface, supports the hypothesis that descending particles in solid phase actively deplete the charge storage of the convective column.

Since pseudo-RHI cross-sections are reconstructed from a sequence of PPI sweeps that require several minutes to complete, the rapid horizontal translation of the storm during this acquisition period can artificially displace upper-level echoes forward relative to lower-level scans. This geometric artifact may induce an apparent tilt in the convective structure, as clearly visible in Figure 4.21.

4.3.3 Multi-Parameter Characterization of Convective Storms

To quantitatively track the convective core evolution directly above the disdrometer location, specific key parameters are extracted from the vertical profiles of horizontal reflectivity (Z_H) and differential reflectivity (Z_{DR}), which are reconstructed following the methodology detailed in Section 3.3. The parameters include the maximum height of the $Z_H > 45$ dBZ region, the thickness of the $Z > 35$ dBZ layer, and the maximum reflectivity values, along with their corresponding height and the associated Z_{DR} measurements. These parameters are synthesized into a three-panel plot to illustrate their temporal evolution. The top panel depicts the vertical structural dynamics, specifically tracking the echo tops and layer thickness, both expressed in km; the middle panel displays the intensity and microphysical state of the convective core through Z_H and Z_{DR} ; and the bottom panel presents the electrical activity, with IC and CG lightning counts synchronized to the radar scan windows as described in Section 3.3. To provide

a comprehensive framework for the event, the temporal axis is segmented using vertical dashed lines representing the specific phases of the event - *Pre-Hail* (cyan line), *During-Hail* (black line) and *Post-Hail* (dark-red line) - as defined in Section 4.2.2 based on the disdrometer measurements. These markers act as a temporal bridge, aligning the atmospheric dynamics captured by the radar and lightning datasets with the surface observations of solid precipitation measured by the ground sensor.

Ev. 08 - June 8th, 2021.

Figure 4.23 captures the temporal progression of radar-derived parameters and synchronized lightning during the June 8th, 2021 storm.

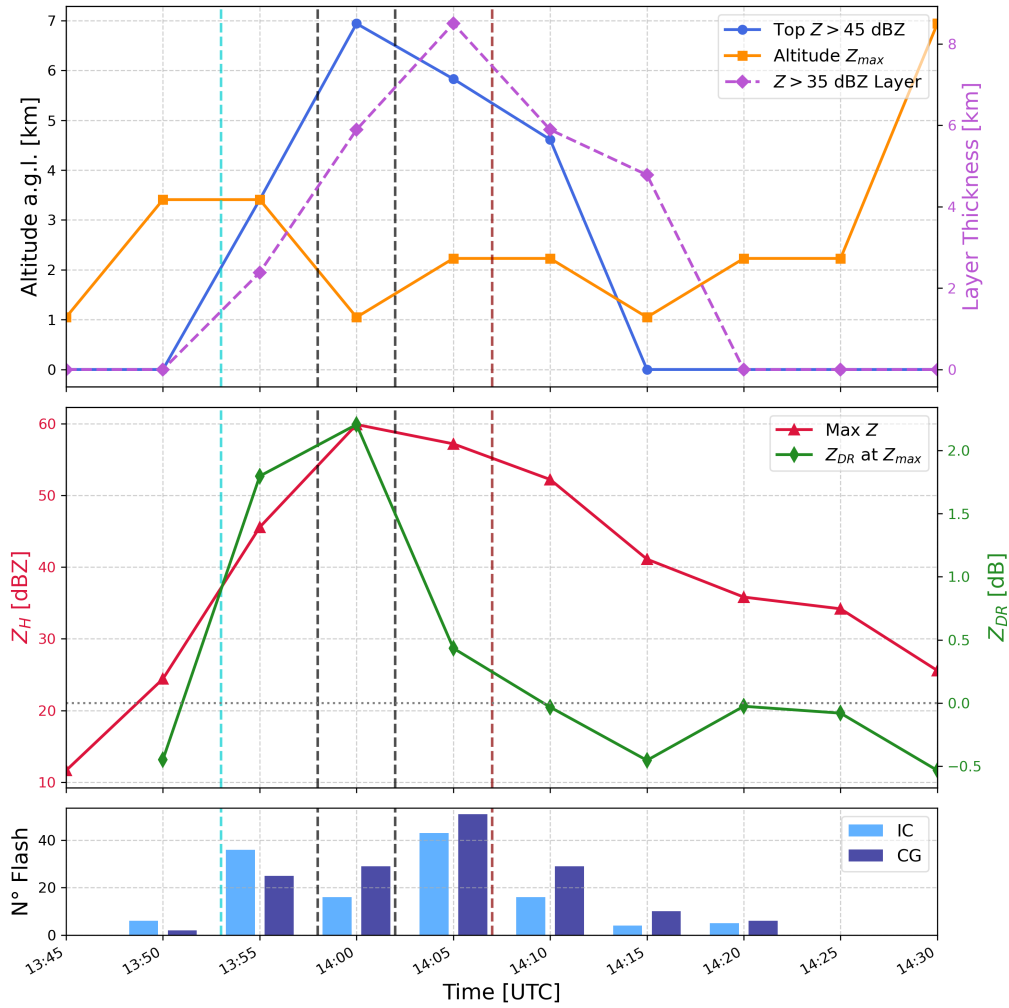


Figure 4.23: Temporal evolution of radar-derived parameters and synchronized lightning activity for Ev. 08 (June 8, 2021). Top panel: vertical structural parameters; middle panel: microphysical variables (Z_H and Z_{DR}); bottom panel: IC and CG lightning counts. Vertical dashed lines delineate the *Pre-Hail* (cyan), *During-Hail* (black), and *Post-Hail* (dark-red) disdrometer-derived storm phases.

In the lead-up to the hail fall, a signature of convective intensification is observed. At 13:55 UTC, the maximum Z_H climbs to 45 dBZ, with the convective core centered at a

mid-level altitude of around 3.4 km. Concurrently, the Z_{DR} rises to 1.7 dB, indicating the presence of growing hydrometeors being lifted by the developing updraft. This structural growth is accompanied by an early surge in electrical activity, in both IC and CG strokes.

The actual surface passage of the hail core is bounded by the vertical black dashed lines. At 14:00 UTC, the altitude of maximum reflectivity drops to its lowest level at 1.1 km, while Z_H hits its absolute peak at 60 dBZ. Simultaneously, Z_{DR} spikes to 2.2 dB at this exact low altitude. This specific pairing of high reflectivity and positive differential reflectivity close to the surface provides a clear polarimetric signature of a melting hail matrix mixed with heavy rainfall reaching the ground. Interestingly, a trend opposite to other events analyzed in this work is observed, as the upper-level structure continues its vertical extension, with the echo top height rising to its maximum of 7 km at 14:00 UTC.

As the system transitions past the second black dashed line into the *Post-Hail window*, at 14:05 UTC, while the surface hail fall has ceased, the $Z_H > 35$ dBZ layer thickness reaches its absolute maximum at 8 km, indicating that the upper levels of the convective cell are still developed. This delayed structural peak drives the absolute maximum of electrical activity at 14:05 UTC, where the system records its highest flash counts of approximately 43 IC flashes and a dominating 50 CG strokes. Given that this convective event develops as a linear squall line, the electrical surge may well be amplified by lightning contamination from adjacent cells along the convective front of the system. From a microphysical perspective, a progressive transition is observed aloft within the radar column at an altitude of 2.2 km. At 14:05 UTC, the radar core remains highly intense with a Z_{max} of 57 dBZ paired with a decreased Z_{DR} of 0.4 dB. This ice-phase signature further matures by 14:10 UTC, where Z_{max} stabilizes around 52 dBZ and Z_{DR} drops near 0 dB. Throughout this 14:05 – 14:10 UTC interval, a discrepancy emerges between these ice-phase characteristics tracked aloft and the surface disdrometer, which records exclusively liquid precipitation with a maximum drop size of 5 mm. This mismatch could be attributed to the instrument’s sorting algorithm misclassifying melting ice as raindrops, as described in Section 4.1.1, or alternatively, it is possible that the falling hydrometeors escape the ground sensor’s limited sampling area due to strong horizontal winds.

Finally, between 14:15 UTC and 14:20 UTC, the intense convective structure aloft recedes, with the echo top height falling to 0 km, while the $Z_H > 35$ dBZ layer thickness undergoes a sharp reduction. Concurrently, the core intensity decays as Z_{max} decreases down to 35 dBZ, consistent with the weaker and more uniform precipitation fields typical of the stratiform zone and detected by the disdrometer (Figure 4.8). In synchronization with this structural transition, the electrical activity decrease, with both IC and CG flash counts dropping to near-zero values as the sampling site transitions from the leading convective core into the trailing stratiform region of the squall line.

Ev. 12 - August 27th, 2024.

Figure 4.24 illustrates the temporal evolution of the extracted radar parameters for the August 27th, 2024 event. Between 12:20 and 12:35 UTC, the extreme altitude of the echo top, reaching nearly 12 km, confirms the presence of a strong updraft, driving an exceptionally deep vertical cloud development, consistent with previous observations. The combination of high maximum reflectivity (Z_H exceeding 50 dBZ) and near-zero

Z_{DR} at an altitude of approximately 8 km indicates a massive accumulation of mixed-phase hydrometeors aloft during the *Pre-Hail* phase. This suspended volume of tumbling ice particles acts as a charge generator, as confirmed by the synchronized lightning data. Specifically, at the peak of this vertical extension (12:35 UTC), the dataset reveals a spike in cloud-to-ground strokes (exceeding 60 flashes), coinciding with the maximum count of intra-cloud flashes recorded for this system.

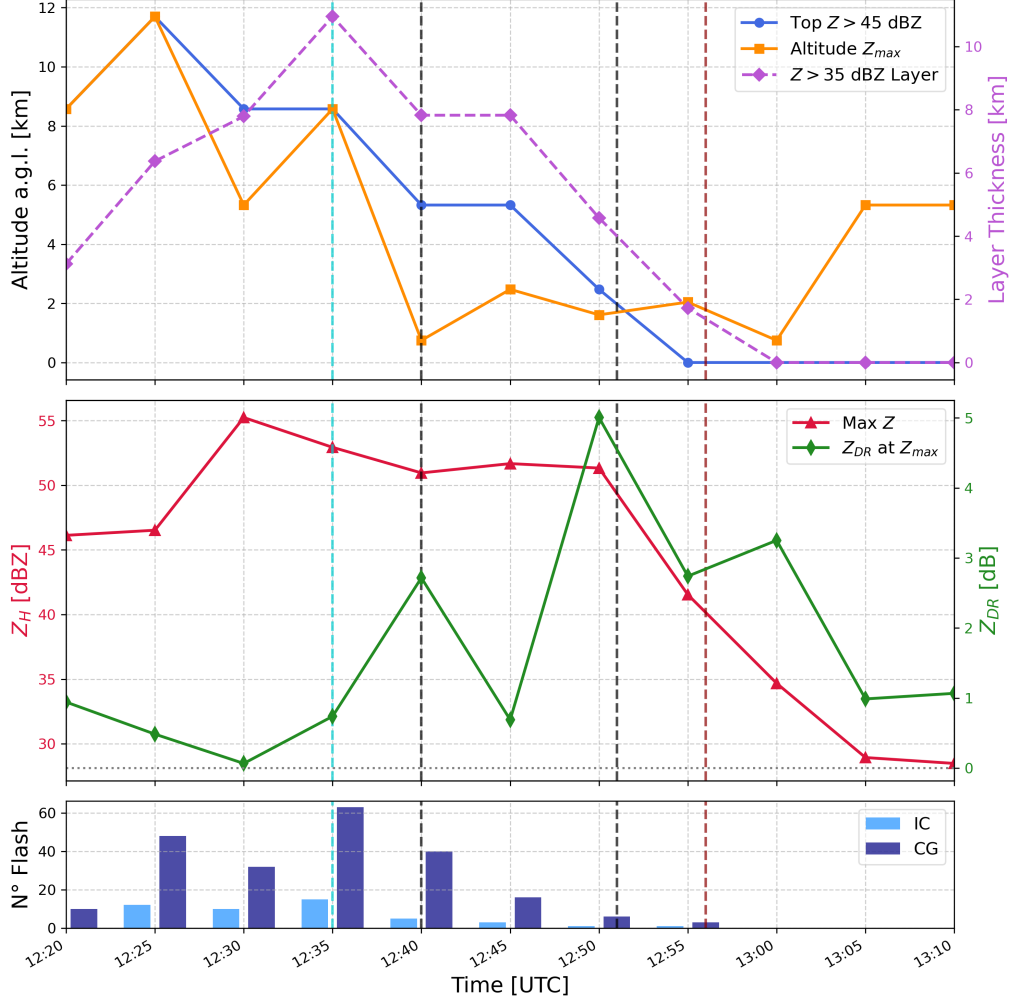


Figure 4.24: Temporal evolution of radar-derived parameters and synchronized lightning activity for Ev. 12 (August 27, 2024). Top panel: vertical structural parameters; middle panel: microphysical variables (Z_H and Z_{DR}); bottom panel: IC and CG lightning counts. Vertical dashed lines delineate the *Pre-Hail* (cyan), *During-Hail* (black), and *Post-Hail* (dark-red) disdrometer-derived storm phases.

At 12:40 UTC, the onset of the hail fall, as captured by the disdrometer, is perfectly synchronized with a drastic structural collapse of the convective core. The altitude of Z_{max} fall down abruptly from 8.5 km down to approximately 1 km. Throughout this descent, the maximum reflectivity maintains high values (above 50 dBZ), but is coupled with a quasi-null Z_{DR} s. This combined signature is perfectly timed with the surface detection of hail, confirming the fallout solid hydrometeors. A progressive decay in the total lightning activity is observed throughout the *During-Hail* window. This synchro-

nized behavior represents again an observational evidence for the hypothesis that the precipitation of solid hydrometeors acts as a physical vector for downward charge transport, effectively leading to the electrical discharge of the convective cloud.

In the *Post-Hail* phase, the cessation of solid precipitation is immediately followed by the complete flattening of the structural parameters. The echo top altitude and the layer thickness progressively diminish to zero, accompanied by a steady decrease in Z_H values. Concurrently, the microphysical signature exhibits a transition: Z_{DR} increases, peaking at approximately 5 dB. This high differential reflectivity, combined with the relaxing Z_H , indicates a shift to purely liquid precipitation as effectively observed by the disdrometer. Finally, the definitive termination of the electrical activity confirms the total discharge and complete fallout of the local convective core.

Ev. 17 - August 21st, 2025.

Figure 4.25 shows the time series of radar-derived parameters and synchronized lightning during the August 21st, 2025 storm.

During the initial phase (13:55 – 14:05 UTC), the Z_{max} fluctuates between 35 and 43 dBZ, while Z_{DR} values remain below 2 dB. The vertical structural metrics show no significant convective development, which is consistent with the low-intensity precipitation registered at the surface by the disdrometer. Starting from 14:05 UTC, a massive structural development is observed instead, culminating at 14:10 UTC when the echo top height ($Z_H > 45$ dBZ) reaches 8.5 km and the $Z_H > 35$ dBZ layer thickness expands to nearly 8 km. At this precise timestamp, a severe convective core aloft is identified by a maximum horizontal reflectivity exceeding 60 dBZ located at a Z_{max} altitude of approximately 5 km, coupled with a minimum in differential reflectivity ($Z_{DR} \approx 0$ dB). This polarimetric combination at high altitudes represents the signature of dry solid particles, such as graupel or developing hailstones, suspended within the mixed-phase region by a strong updraft.

Following this peak aloft, the core undergoes a rapid descent, with the Z_{max} altitude dropping to 1.6 km. Concurrently, Z_{DR} spikes to approximately 5.3 dB while Z_H sustains its peak at 63 dBZ. This polarimetric signature at low altitudes suggests the likely presence of melting solid hydrometeors entering the lower levels, rather than a purely liquid downdraft. However, at this timestamp (14:15 UTC), the disdrometer registers exclusively liquid precipitation at the surface characterized by exceptionally large raindrops, classified by the sensor in the bin with a mean diameter of 7.5 mm, as visible in (Figure 4.12). This discrepancy could highlight the well-known misclassification bias of the disdrometer’s sorting algorithm, already noted in the previous DSD spectral analyses (Section 4.1.1). Since melting hailstones are enveloped in a liquid film, they are highly susceptible to this misclassification as rain by the ground sensor. The total lightning activity undergoes a general intensification before reaching its absolute peak precisely within the Pre-Hail window at 14:20 UTC, where it records approximately 30 IC flashes and 40 CG strokes.

The actual surface passage of the hail core, bounded by the vertical black dashed lines (14:22 – 14:31 UTC), is characterized by Z_{max} values ranging between 45 and 55 dBZ, stabilizing at the low-levels, and by positive Z_{DR} values fluctuating between 2.6 and 4.8 dB. This polarimetric signature near the surface confirms the arrival of a complex melting hail matrix mixed with heavy rainfall. Simultaneously, the top panel reveals a

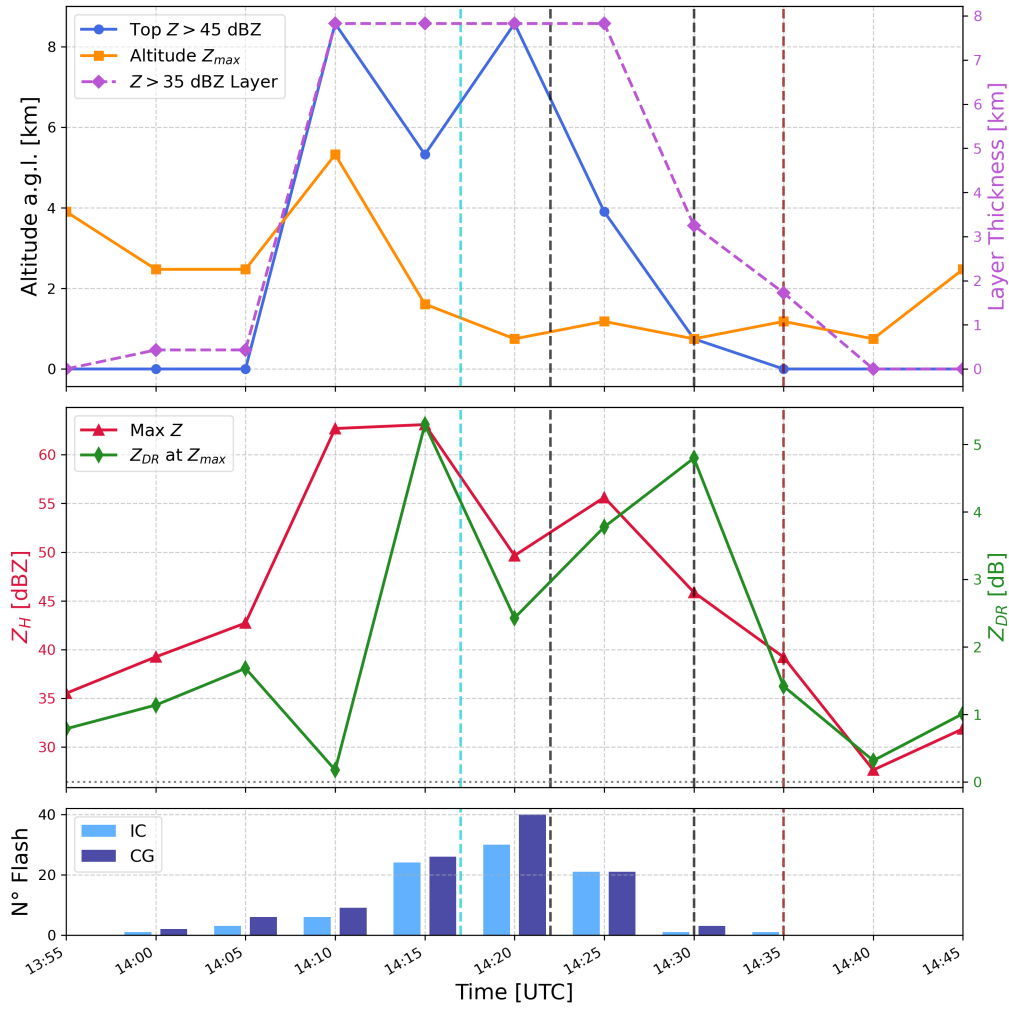


Figure 4.25: Temporal evolution of radar-derived parameters and synchronized lightning activity for Ev. 17 (August 21, 2025). Top panel: vertical structural parameters; middle panel: microphysical variables (Z_H and Z_{DR}); bottom panel: IC and CG lightning counts. Vertical dashed lines delineate the *Pre-Hail* (cyan), *During-Hail* (black), and *Post-Hail* (dark-red) disdrometer-derived storm phases.

rapid structural collapse of the convective cell: between 14:25 and 14:30 UTC, the echo top height ($Z_H > 45$ dBZ) drops from 4 to 0 km, while the $Z > 35$ dBZ layer thickness shrinks by more than half. Driven by this structural collapse and core grounding, the electrical activity undergoes a progressive collapse throughout the *During-Hail* window. By 14:30 UTC, both IC and CG flash counts decay to near-zero values.

Passing the second black line into the *Post-Hail* window (14:31 – 14:36 UTC), the storm enters a definitive weakening stage. The structural parameters collapse toward zero, while Z_H decreases below the convective threshold, dropping under 40 dBZ. Concurrently, the microphysical signature shows a minor adjustment in Z_{DR} and a total cessation of electrical activity is observed.

Chapter 5

Conclusions and Outlooks

This research investigated the relationship between microphysical and electrical dynamics within severe convective cores over the metropolitan area of Bologna. By integrating ground-based disdrometer measurements, polarimetric weather radars, and total lightning data, this study aimed to determine whether solid precipitation acts as a primary downward charge carrier.

The multi-sensor analysis allowed to draw significant insights regarding the electrical response of thunderclouds to solid precipitations.

First, regarding the depolarization process, this study points out that the electrical activity of severe convective cells is modulated by the kinematic evolution of solid hydrometeors. The highly electrified phase of a storm is strictly coupled with the suspension of high-reflectivity ice cores ($Z_H > 50$ dBZ) aloft by intense updrafts, acting as charge reservoirs. The subsequent structural collapse and fallout of these solid matrices lead to a macroscopic decrease of the storm's lightning activity.

The analyzed hail events exhibit an interesting departure from standard regional climatology. Rather than being dominated by IC pulses, these extreme convective cores show a complete inversion in the discharge partition, with CG strokes being dominant (up to 70%) over IC discharges. Furthermore, IC emission heights are shifted toward lower atmospheric levels, providing a distinct electrical signature of descending solid convective cores. Extended research should be conducted to determine whether this correlation is a systematic feature of severe hailfall events or a specific anomaly.

Characterizing these complex dynamics highlighted critical vulnerabilities in ground-sensor networks. The disdrometer measurements of anomalously large hydrometeors, up to 11.0 mm in diameter, within spectra that should contain only liquid particles reveal a systematic cross-contamination between precipitation phases, severely biasing the D_m distribution. Without a proper polarimetric radar validation, the disdrometer's particle sampling proves to be unreliable during mixed-phase precipitation, particularly in the classification of melting hail. Therefore, future efforts should focus on the development and implementation of classification algorithms capable of properly identifying melting hydrometeors.

In addition, this study suggests that single-point ground sensors are insufficient to fully characterize convective systems, as they cannot distinguish whether apparent ground level anomalies, such as delayed electrical peaks after hailfall, temporary precipitation interruptions, or peak in lightning activity preceding the surface precipitation, are due

to actual microphysical and dynamical evolution or are geometric artifacts caused by the system's advection. By mapping the 3-D structure of the storms, this work proves that disdrometric observations require a broader volumetric framework to resolve this ambiguities. For example, a disdrometric network could be deployed to provide a wider spatial footprint of surface precipitation, enabling a more accurate correlation with the storm's electrical response during its entire advection path.

To mitigate the temporal lag introduced by volumetric PPI scans in rapidly advecting systems, future observational strategies should incorporate RHI scans, which provide an almost instantaneous view of the storm's vertical architecture, capturing the precise timing of hydrometeor fallout, thereby improving the understanding of the underlying depolarization processes.

A significant advancement is represented by the implementation of 3-D lightning mapping operating in the VHF band, which would allow for the tracking of the exact volumetric location of electrical discharges in relation to the suspended high-reflectivity ice cores. Combining ground-based measurements with the space-based total lightning observations from the new Lightning Imager (LI) on board the Meteosat Third Generation (MTG) satellites will enable the continuous tracking of severe convective cores throughout their entire lifecycle.

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